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THE KILAUEA EAST RIFT ZONE : GEOTHERMAL EVALUATION OF THE EXISTING DATA

DRAFT REPORT

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ENEL

ENTE NAZIONALE PER L'ENERGIA ELETTRICA - ITALY

LIST OF FIGURES, PLATES, ANNEXES AND TABLES

LIST OF ABBREVIATIONS USED IN THIS REPORT

DBED – Department of Business and Economic Development

DLNR – Department of Land and Natural Resources

DOE – Department of Energy

GCP – Geothermal/Cable Project

GRS – Geothermal Resource Subzones

ERZ – East Rift Zone

HECO – Hawaiian Electric Company Inc.

HGP – Hawaiian Geothermal Project

HIG – Hawaiian Institute of Geophysics

HVO – Hawaiian Volcano Observatory

KERZ – Kilauea East Rift Zone

LERZ – Lower East Rift Zone

MERZ – Middle East Rift Zone

SOH – Scientific Observation Holes

TMP – True/Mid-Pacific

USGS – United States Geological Survey

THE KILAUEA EAST RIFT ZONE: GEOTHERMAL EVALUATION OF THE EXISTING DATA

1. INTRODUCTION

Approximately 90% of the electric power of the Hawaiian Islands, which are isolated in the middle of the Pacific Ocean, is produced from imported oil due to the lack of traditional energy resources such as coal and large hydropower facilities.

The almost total dependency of the State of Hawaii on imported oil places it in an extremely vulnerable position in the event of an oil crisis such as that of the 1970's.

For this reason, various research and development projects have been launched in recent years in order to analyze the possibility of exploiting indigenous resources such as solar, wind, biomass, small hydropower, ocean thermal and geothermal energy.

At present, however, only geothermal energy can make a significant contribution within a short period of time to the production of electricity and hence diminish the State's great dependency on fuel oil.

Geothermal resources are, in fact, present on the Island of Hawaii which is the largest and youngest of the islands in the Hawaiian chain and where the active volcanoes Mauna Loa and Kilauea are located. On this island the Kilauea East Rift Zone (KERZ) was selected as the most promising site for the geothermal exploration.

The first exploratory wells were drilled on the KERZ area in 1961 and 1962. Four shallow wells were completed to depths ranging from approximately 54 m to 210 m; the maximum temperature measured was of about 100°C. They were subsequently abandoned because none of these wells was considered suitable for electric production.

In 1972 the Hawaii Geothermal Project (HGP) was initiated by the University of Hawaii College of Engineering in order to verify the presence of an industrially exploitable geothermal system on the KERZ area. Geological, geochemical and geophysical surveys were performed and in 1975 the drilling of the first exploratory deep well, called HGP-A, was initiated.

The well was productive and demonstrated the presence of a high temperature hydrothermal system. In the period from 1981 to 1989 this well supplied a 3 MW condensing unit and confirmed that the resource was usable for electric production.

Other 7 deep exploratory wells were drilled in the period from 1980 to 1985 by two private companies in areas close to the HGP-A well and three of them were found to be productive.

On the basis of the data available, the power generating potential of the KERZ area has been estimated by some scientists to be in excess of 1000 Megawatts for the next century.

The State of Hawaii is investigating the possibility of developing the geothermal resource available in the KERZ area and to this extent the Geothermal/Cable Project (GCP) was outlined. The planning includes well drillings and the construction of surface equipment and power plants for a total installed capacity of 600 MW.

The Island of Hawaii will be supplied with 50 to 100 MW to cover the estimated increase in its energy requirements over the coming 20-30 years (to date, the installed capacity on the island is only 101 MW). Five hundred MW will be for the Island of Oahu (Honolulu), where the presently installed capacity is 1255 MW to meet the foreseen increasing demand.

The electric energy will be transported from Hawaii to Oahu through an interisland transmission system including submarine cables. It is envisaged that three submarine cables, each carrying a total power of 250 MW, will be laid for this purpose.

The deep water cable system will cover a distance of around 240 kilometers between the Island of Hawaii and the Island of Oahu. The cables will be laid at a depth of about 1900 meters (the greatest depth at which an underwater cable has been laid to date) in the sea channel that separates Hawaii from Maui islands.

Before initiating the project development, which implies large capital investments, it is necessary first of all to verify the presence of geothermal resource in the area selected for the project, to evaluate the field potential and to define the chemical-physical characteristics of the fluids produced. It is necessary, moreover, to verify the feasibility of the electric power transfer for such great distances and depths by means of submarine cables.

Researches and tests on submarine cables have been performed in the frame of the Hawaii Deep Water Cable Program. In 1985, 1800 meters of cable were manufactured for the laboratory tests and in 1988 the surveys on the oceanic floor and the laboratory tests were both completed. A surrogate cable was tested at sea in 1989.

The results of the tests confirmed that the installation of a deep water cable system between the Hawaii and Oahu islands is technically feasible.

The presence of the geothermal resource in the entire area selected for the project, instead, is still to be verified. A Geothermal Resource Verification and Characterization Program was drawn up envisaging first the drilling of four Scientific Observation Holes (SOH) and subsequently the drilling of about 25 deep commercial exploratory wells together with the performance of measurements and tests.

Data and information collected from this program are fundamental for the project and strongly influence the development phase of the Geothermal/Cable Project.

On June 1990 only one SOH and a deep exploratory well were completed. Data related to these two wells are not yet available.

The Geothermal/Cable Project will be developed by a consortium of private companies that will be chosen by the Hawaiian Electric Company Inc. (HECO), which will purchase electric energy for Oahu.

Requests for proposals were distributed by HECO in May 1989 and five consortia, which constitute major companies worldwide, responded. The technical proposals were received in November 1989, and the economic proposals in December 1989. In February 1990 HECO selected two consortia with which negotiations are ongoing in order to make the final choice by the end of 1990.

According to HECO's request for proposals, the first 125 MW should be available for Oahu by 1995.

2. OBJECTIVES

Following contacts initiated in 1989, the Government of Hawaii has decided to avail itself of ENEL as a consultant on some of the activities concerning the development of geothermal resource in the KERZ area.

The economic dimensions of this project and its strategic, political and social significance demand the involvement of the Government of Hawaii which has allocated funds destined to the Geothermal Resource Verification and Characterization Program.

The estimates of the field potential carried out to date for the KERZ area are based on a series of hypotheses that in great part must still be verified.

At present, even if in this area high temperatures can be reasonably assumed, there is no information about the permeability distribution and the chemical-physical characteristics of the fluids inside the reservoir. It is therefore necessary to plan and carry out a program of activities aimed at assessing the existence of an economically and industrially exploitable geothermal resource.

The objectives of this work are:

- to review and analyze the activities and studies carried out to date for the geothermal characterization of the KERZ area;
- to provide, on the basis of these analyses, a series of indications and recommendations for the Geothermal Resource Verification and Characterization Program.

3. CHRONOLOGY, PARTICIPANTS AND ACKNOWLEDGEMENTS

The general advising, the technical documents collection and the meetings with the experts involved in geothermal activities initiated in 1989.
The list of the Italian experts and the mission periods to Hawaii regarding the activities related to this report are given here below:

Name	Qualification	Period	
Guido CAPPETTI	General Adviser	2-10 Sept.	1989
		10 Nov.-16 Dec.	1989
		2-12 Jan.	1990
		5-29 June	1990
Raul TONEATTI	General Adviser	2-10 Sept.	1989
Giorgio BUONASORTE	Geologist	10-25 Nov.	1989
		15-29 June	1990
Adolfo FIORDELISI	Geophysicist	10-25 Nov.	1989
		15-29 June	1990
Gianni SCANDIFFIO	Geochemist	10-25 Nov.	1989
		15-29 June	1990

In addition to the personnel above, Mauro CAMELI (Geophysicist) and several other ENEL's experts have collaborated in Italy for the data analyses and processing.

A major role was played by the Department of Business and Economic Development (DBED), which made available a lot of technical documentation and organized the meetings. We would like to particularly thank Mr. Ulveling, Mr. Kaya and Mr. Lesperance for their active collaboration.

We would also like to thank the experts with whom, in the course of various meetings, aspects of the geothermal problems in Hawaii were discussed, and who offered useful suggestions.

Amongst them, in particular: A. Furumoto, C. Helsley, J. Kauahikaua, H. Olson, D. Thomas and G. Walker.

Thanks go finally to the representatives of the Department of Land and Natural Resources (DLNR) and of the two private companies (TRUE/MID PACIFIC and ORMAT) with whom several meetings have been held.

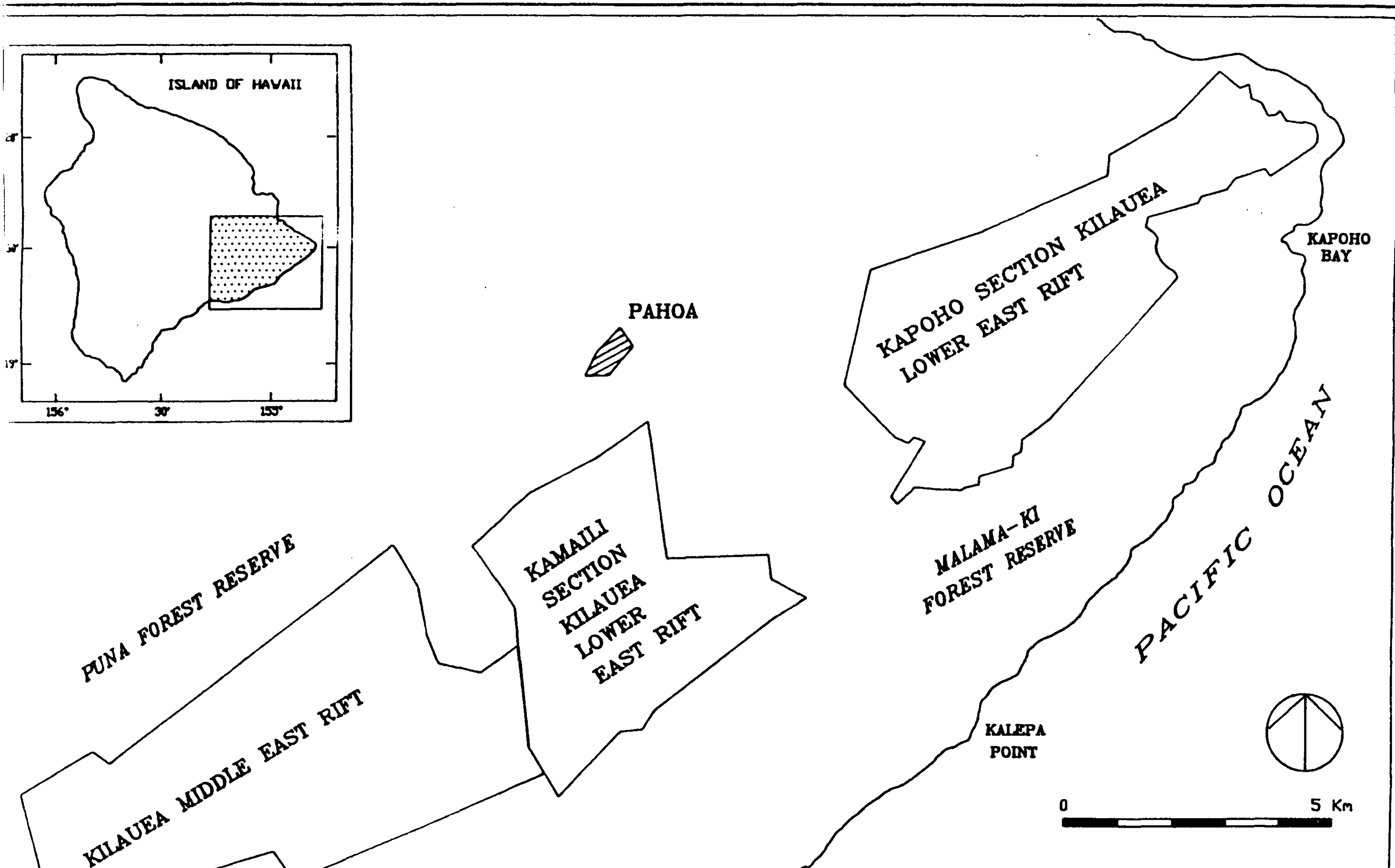

Figure A
GEOTHERMAL RESOURCE SUBZONES

4. SETTING OF THE PROJECT AREA

The Kilauea East Rift Zone, on the Island of Hawaii, has been characterized, in historic times, by frequent volcanic activity.

Geological, geophysical and geochemical surveys were carried out by several State and Federal Agencies (Universities, Hawaiian Volcanoes Observatory, Department of Energy, etc.) mainly focused on the general geologic and volcanologic studies. More detailed surveys have been recently carried out and the deep geothermal wells drilled in the Lower East Rift Zone have confirmed the presence of high temperatures and of an industrially exploitable hydrothermal system.

In this area the Department of Land and Natural Resources (DLNR) has selected and defined three Geothermal Resource Subzones (GRS), on the basis of the existing environmental constraints such as national parks, human settlements, rain forest, etc. All the activities relative to the project (wells, surface equipment and plants, power stations) will be developed within these Subzones which are located in the easternmost part of the rift and are called KILAUEA MIDDLE EAST (9014 acres - 36.47 km^2), KAMAILI (5530 acres - 22.38 km^2) and KAPOHO (7350 acres - 29.74 km^2), covering a total area of 22000 acres - 88.59 km^2 (Fig.A).

5. ANALYSIS OF THE EXISTING DATA

There are many published papers and reports from scientific and private organizations, including data about performed drillings and have been gathered only partially and in a qualitatively dishomogeneous way.

More than 150 documents have been analyzed synthesizing the most important data for a geothermal reconstruction both of the regional structure and of the geothermal system model.

This report therefore represents an up-to-date and synthetic review of the available data. All data are assembled and organized in thematic Plates and Annexes; in particular, Annex A refers to the Geochemistry data, Annex B refers to the Well data. In Annex C there is the complete list of the collected and analyzed documents.

5.1. GEOLOGY

Kilauea is certainly the most studied volcano in the world. The characteristics of its geology, volcanology, petrography and tectonic-structural setting are described in numerous papers. In this report only the aspects relevant to a geothermal interpretation of the eastern side of the volcano are considered. Kilauea, which resembles Mauna Loa Volcano both in structure and evolution, is one of the youngest volcanoes in the Hawaiian Island Archipelago. Kilauea has grown by over 6500 m (1200 above sea level) from the ocean bottom on the eastern flank of Mauna Loa.

5.1.1. THE KILAUEA VOLCANO.

Structural outline

The Kilauea volcano can be divided into 6 volcano-tectonic areas (Fig. 1/P1) (Holcomb, 1987).

Summit. It is characterized by an annular collapse structure intersected by SW-NE fault systems. The present caldera is very young (age < 500 years). It is 5 km long, 3 km wide and about 120 m deep. The principal dislocations of the present caldera go back to the 1790 eruption.

The caldera has developed in the same area where probably there had previously been another one (Powers Caldera).

Small pits such as Halemaumau, Kilauea Iki and KeanaKakoi crater also occur. The eruptive activity has been effusive, of persistent character, and in lesser extent explosive.

Koae Fault System. South of the summit area intense SW-NE fracturing occurs. The north downthrow faults might constitute the buried margin of an older caldera. Of the few known eruptions, the greater part occur at the intersection with the East Rift structure.

South-West Rift Zone. It develops along 32 km parallel to a similar rift zone of the Manua Loa. A migration of the volcanic activity towards the sea can be deduced from the age of the eruption vents. The most highly differentiated products (viscous lavas and scoria cones) in all Hawaii have erupted here.

East Rift Zone. This is a distensive structure, the surface width of which varies from 4 to 6 km. Its length is 45 km and its submarine continuation is over 70 km long. It is divided into an upper, a middle and a lower part. The upper part is dominated by pit craters with essentially effusive activity of modest duration.

The middle and lower parts have been affected by brief and numerous eruptions in which the production of pyroclastic cones was frequent.

South Flank. The southern flank of the Kilauea volcanic system is located between the Koae Fault System, the East Rift zone and the sea. Northward lowering faults occur only occasionally. The external zone is dissected by southwards downthrow faults of the Hilina System. Probably in the internal part the faults were covered by lava and never again reactivated. The age of the System is unknown. The fault system might pre-date Kilauea. Their total displacement is of at least 300 m estimated on the basis of field outcrops (Peterson and Moore, 1987). No eruptions have been registered in this area.

North Flank. There are no surface volcanic structures. It is possible that currently inactive structures have been covered and hidden. There have been no eruptions in this area in the last millennia. Lavas, mostly tube-fed flows, from the summit area, has accumulated here.

Stratigraphy

Kilauea is formed by a sequence of basaltic submarine lavas thinly stratified and deposited at depths that became increasingly smaller as the volcano grew.

Only three units have been identified at the surface. From the oldest to the most recent, they are: **Hilina Basalt, Pahala Ash and Puna Basalt** (Langenheim, Clague, 1987; Easton, 1987).

The Hilina Basalts are over 25000 years old. They are exposed over modest areas along the slope of the Hilina Fault System with a maximum thickness of 300 m. They are tholeiitic, olivin-tholeiitic and picritic basalts to which pyroclastic palagonitic deposits and ash intercalations often palagonitized are associated with thickness varying from a few centimeters to several meters. The latter are from the bottom: Halape Ash Member, Hahele Ash Member, Pohaka Ash Member and Moo Ash Member.

The Pahala Ash is an intensely palagonitized vitric ash common to the stratigraphy of both Kilauea and Mauna Loa. Along with Moo Ash and Pohaka Ash, with which it shares chemism and distribution, the Pahala Ash might be the product of the violent, explosive eruption of either Mauna Loa or Kilauea. In the area of Kilauea, the Pahala Ash occurs only at the top of the few outcrops of Hilina Basalts for a maximum thickness of 15 m. These units are between 10000 and

25000 years old.

The Puna Basalts are the result of long sustained eruptions and lateral eruptions in the rift zones which occurred between 10000 and 1500 years ago. They cover almost the whole surface of the volcano, which for its greater part (90%) is younger than 1100 years (70% less than 500 years). They are essentially tholeiitic basalt and picritic tholeiitic basalt flows. The maximum exposed thickness is of about 100 m. Minor pyroclastic intercalations are present. They are ashes and pumices of the pyroclastic surge associated to 1100-1500 years old and 1790-1823 summit phreatomagmatic eruptions (Uwekahuna and Keanakakoi Ash Members respectively).

The old Powers caldera, whose collapse and filling up occurred between 1500 and 1100 years ago, was formed at the same time as the deposition of Uwekahuna Ash Members. This period is characterized by an essentially intracalderic activity.

Widespread calderic overflows and rift eruptions, without any evident link among them, occurred up to 350 years ago.

Thereafter summit activity has been persistent. It has been followed by rift eruptions which, by draining the main conduit, have caused one of the most important collapse events of the present caldera. The phreatomagmatic explosion of 1790, from which Keanakakoi Ash originated, accompanied that collapse event.

From 1790 and for over a century all the surface activity was intracalderic, almost always persistent, with the exception of two lateral eruptions in 1823 and 1840. This activity stopped in 1924 with a phreatomagmatic explosion (Halemaumau Crater), which was perhaps followed by eruptions in the rift which coincided with the summit collapse.

Since then summit eruptions have been sporadic and brief, while rift eruptions have become more frequent and persistent (Holcomb, 1987).

5.1.2. SURFACE STRUCTURAL SETTING OF THE KERZ

The main volcanic and structural features of the Kilauea Summit and its East Rift are summarized in Plate 1. In the absence of a published geological map of an adequate scale, the main features were redrawn schematically from the Holcomb's (1980 and 1987) Preliminary and Reconnaissance Geological Maps slightly modified on the basis of the data provided by other sources. The symbols do not always correspond to the real extension of the features.

The KERZ develops southeastwardly as far as Mauna Ulu, where it intersects with the Kaoe Fault and then it changes towards the East North East in the direction of Cape Kumukahi.

The Upper East Rift Zone runs from the summit caldera eastwards as far as the Napau Crater to an altitude of approximately 800 m. The KERZ is obliquely cut by WSW-ENE en echelon eruptive fissures. It also presents several volcanic structures, active in prehistoric and historic times, such as cones, pit craters and fissure vents. Among these, the 500 years old Puu Huluhulu spatter cone, and the Mauna Ulu, Alae and Kane Nui O Namo lava shields stand out from a morphologic point of view. More recently, the following centers have been active: 1922 Makaopuhi, Napau; 1961 Napau; 1963 Alae Crater; 1963 Alae and Napau; 1965 Makaopuhi, Aloi; 1968 Hiiaka, Kane Nui O Namo; 1969 and 1972 Mauna Ulu, Alae, Aloi and Napau.

The Middle East Rift Zone continues toward East North East until the Heiheiahulu satellite Shield. This area has few pit craters; however, it presents many eruption vents. This part of the rift is characterized by swarms of numerous faults and fissures which sometime form small graben with a large longitudinal development. An important volcanic activity occurred especially in the southern area of the rift. In 1965 there were eruptions from the Kalalua vent. The Puu Oo lava shield started to grow in 1983; this is the largest vent edifice of Kilauea with a diameter of 1 km and an altitude of 255 m over the preexisting surface. The summit of Kilauea has been subject to deflation phenomena during each of the eruption episodes of Puu Oo, whereas during the intervals it has displayed inflating episodes. The chemism of the undifferentiated lavas emitted is identical in all its characteristics to that of the summit lava. This suggests the existence of a direct link between the summit magmatic reservoir and the area of Puu Oo. In 1986, after a lava fountain activity along a fissure of about 5 km, the eruption concentrated approximately at the present Kupaianaha vent (Hazlett, 1987).

The Lower East Rift Zone extends from an elevation of approximately 400 m to the sea. It is also characterized by a few pit craters and several volcanic edifices. The latter are frequently explosive because of the interaction of the magma with groundwater (meteoric or marine). The rift is formed by a wide ridge with a weak longitudinal slope which meets the sea at Cape Kumukahi. There are fissure vents of considerable length in this rift zone. Fewer faults and fractures can be seen especially in the area south of Pahoa. In 1790 many structures in this part of the rift were buried by voluminous flows fed by centers in Heiheiahulu, Puu Kaliu and other scattered vents (Moore, 1983). The lava flowed with a weak slope towards the north and especially the south. There are prehistoric cinder and spatter cones (Puu Kukii, Puu Kaliu, Puulena - Honuaula) and tuff rings of phreatomagmatic origin (Kapoho Cone). More recently, in 1955, there were voluminous eruptions along a 14 km long series of fissures that developed immediately to the east of Heiheiahulu Cone to the area of the Kapoho village. A new eruption occurred in 1960 which built the new scoria cone of Puu Laimama and emitted large lava flows.

The two eruptions first fed tholeiitic lava flows; then their chemism became increasingly similar to the olivine-controlled summit chemism (MacDonald and Eaton, 1955; Bullard, 1978).

5.1.3. MAGMATIC CHAMBERS AND INTERNAL STRUCTURES OF THE KERZ

The magmatic chamber of the Kilauea should be located at a neutral buoyancy depth, between the shallower and less dense lava flows and the older lava flows which are denser because of magma injections (dikes or intrusions). In this form the magma would be gravitationally stable (Walker 1986, 1988).

The inflation periods which characterize the magmatic activity of the KERZ suggest storage in the chamber. Rapid deflations, instead indicate migrations of a highly mobile magma in the plumbing system connecting the different parts of the volcano. Such migrations are linked to eruptions or injections which give rise to the Dike Complex of the KERZ.

Petrochemical data suggest the existence of a main central reservoir and of secondary satellite reservoirs along the KERZ (Fig. 2/P1). On the basis of geophysical data the depth of these reservoirs is estimated at between 2 and 6 km. The Chain of Craters area would be characterized by large shallow magma intrusions (Macdonald and others 1983). In the same way, Walker (1988) suggests that these intrusions reach the surface because of the collapse of the dike vaults which eventually generate the pit craters through summit collapse.

All over the rift, and particularly near Heiheiahulu Vent, Kaliu Vent and Puulena-Honuaula Vents, where the presence of magmatic chambers has been suggested, there are great quantities of differentiated magmas (Moore, 1983 and related references).

With few and rare exceptions such as the 1959 eruption, magmas feeding the central or lateral eruptions seem to pass through the central magmatic chamber. (Wright and Fiske, 1971; Wright and Helz, 1987).

It is believed that summit eruptions are fed from the highest part of the central reservoir, while the rift eruptions are fed from the deepest part of the rift itself, where a compositional layering would determine an olivine increase. During eruptions or intrusions in the rift, quantities of magma are displaced and settled, accompanied by cooling and, for longer storage periods, differentiation phenomena occur through fractioned crystallization processes (Wright and Fiske, 1971; Wright and Helz, 1987).

In addition, mixing processes, involving the new and residual magma left from previous intrusions, occur in the reservoirs of the rift.

The historical lava flows of the rift originate: from tholeiitic magmas differentiated in the chambers, from olivine controlled magmas (coming directly from the central reservoir) and, finally, from hybrid mixing derived ones. Non modified lavas, similar to the summit (olivine controlled) lavas, are less common in the rift than other lavas.

The erosion and exposure of the Koolau lava shield volcano and its intrusive Dike Complex on Oahu to their inferred depth of 1000 m allowed the study of the internal structures of the Hawaiian rifts. It is thought that their structural location and their evolution are very similar to those of Kilauea (Walker, 1986, 1987). The rift is injected by dikes with an average width of 65 cm and a rate of occurrence between 50 and 100 dikes per 100 m. These dikes frequently tend to intrude on the borders of those previously existing and are associated in clusters up to 20 m wide. The dikes are oriented in almost the same direction as the rift; they form two complementary sets with opposite slopes of 65° to 85°. The lowest frequency of the dikes occurs towards the borders of the rift and in the higher parts which, being younger, are affected only by the more recent intrusions. The low viscosity of the magma and the availability of many access paths (contraction joint, dike margin, flow unit boundary, rubble layers etc.) allow the movement of the magma and its intrusion as dikes, conditioned by the stress field. In turn this phenomenon is determined by hydrostatic forces, which allow the magma to settle in those areas where it is more stable from a gravitational point of view. The dikes do not present a Gaussian distribution but rather an average of

50-65 dikes/100 m. This frequency is fairly constant in the rift; only locally it exceeds 80%. In consequence the zoning of the bulk densities is as follows: lava flow unit 2.3 g/cm^3; dike complex 2.7 with 50-65% in dike volume (dike density 3 g/cm^3); marginal part of the rift with intermediate density (Walker, 1986).

Assuming that each fissure eruption marks a dike, the number of injections in the proximity of the crater of Kilauea can be estimated at 28 injections in 200 years, without taking into consideration the non eruptive injections which may have occurred. It is very unlikely that each magma batch produces a new dike. In any case they feed the satellite magmatic reservoirs in the rift.

The number of dikes decreases with an approximately logarithmic distribution, as the distance from the central part of the volcano increases. In the northeastern rift of Mauna Loa the decrease is greater probably because it is bordered to the north and to the south by the edifices of Mauna Kea and Kilauea (Walker, 1986). In contrast, the KERZ developed principally lengthwise and its southern flank is free from constraints and is highly mobile (Walker, 1988). In fact, the intrusive force of the magma is the cause of the southward displacement of the active part. The intrusion at depth and emission at the surface of enormous magmatic masses would generate gravitational stresses on the growing flank of the volcanic edifice. This would create a set of blocks which, separated by southwards downthrow faults would constitute a kind of megaslides. In this way the southern slope of Kilauea would be in constant subsidence (Walker, 1988; Hazlett, 1987; Macdonald and others, 1983,).

The lateral expansion of the rift and the displacement of the South Flank can be reasonably assumed to occur with deep listric faults, or with horizontal detachments or with side slip movements (Swanson and others, 1976; Walker, 1988). The structural deformations linked to the magma intrusion result in the tectonic collapse of the unbuttressed southern flank towards the sea. In turn, such tectonic phenomena can trigger volcanic activity in the satellite magmatic reservoirs.

Large scale re-mobilizing and mixing of new and preexisting magma are believed to derive from strong earthquakes which may disconnect the conduits. This originates chemical changes in the eruptions that cannot be attributed only to magmatic processes but are closely connected to volcanotectonic constraints (Wright and Helz, 1987 and related references).

5.2. HYDROGEOLOGY

5.2.1. TEMPERATURE AND RAINFALL

The temperature and precipitation distributions are conditioned by the topography. The annual average temperatures in the coastal areas range between 22 and 24°C; it diminishes with height, with smaller gradients in the leeward areas (Thomas and others, 1979). The abundant rainfalls which characterize most of the Islands of Hawaii are a consequence of the interaction between the topography and the masses of humid air of the trade winds that blow from the NE (Feldman & Siegel, 1980). Generally, then, the NE slopes, receive the greater rainfalls; in some locations, such as the Halekala, Mauna Loa and Mauna Kea, the precipitation may be more than 3500 mm/year. In contrast, rainfall in the more arid leeward areas displays values of a few hundreds of mm. In the area of the KERZ the average rainfall is of approximately 2500-3000 mm/year (Fig.1/P2) (Taliaferro, 1959; Thomas and others, 1979).

5.2.2. HYDROGEOLOGICAL CHARACTERISTICS OF THE OUTCROPPING ROCKS

The area under study is characterized by few, discontinuous, thin and, in general, low permeability ash intercalations. The basaltic flows have good permeability with numerous infiltration paths such as vesiculated layers flow unit boundaries, rubble layers, contraction joints and lava tubes. In particular, the lava tubes, which may reach a length of more than 1 Km and a width of approximately 15 m, are very high permeability preferential paths. In this way the KERZ and the bordering areas, characterized by a thin soil cover, present high permeability and high infiltration and, consequently, they lack permanent streams. In these conditions rainfall infiltrates rapidly in all areas to reach the basal water (Stearns, Macdonald, 1946; Druecker, Fan, 1976).

In the submarine lava flows, which are gradually less vesiculated as the head of the original ocean water increases, permeability decreases with depth. These lava flows present very low permeability, only a few millidarcy in contrast with the 4000-5000 darcy of the shallow lavas. Hydrothermal argillification and mineralization phenomena contribute to reduce permeability at depth (Druecker, Fan, 1976).

5.2.3. HYDROGEOLOGICAL CHARACTERISTICS OF SHALLOW WATERS

Springs (pools and ponds) linked to the base water occur only along the coastline. In most cases, they contain brackish waters because of tidal mixings (Thomas and others, 1979). Hence, the data on shallow waters derive from wells, the depths of which vary from a few to several hundred meters and which were drilled principally for water supply and geothermal measurements. Table 1/P2 shows the main data of the wells, including the deep geothermal ones which will be discussed in detail (Druecker and Fan, 1976; Thomas and others, 1979; Epp and Halunen, 1979; Thomas and others, 1983; Imada, 1984). Almost all the wells are concentrated in a small area around the Lower East Rift; only a few are to the south or further north of the rift. The data shown in the table, including the geographical data, come from diverse bibliographical sources which frequently disagree with each other. The parameters selected and shown in the table should be reliable even if there are some doubts as to the accuracy of some of them. In particular, some coordinates were calculated on the basis of oral communications provided by some authors.

In the absence of impermeable barriers, fresh waters should be present in the island in accordance with the Ghyben-Hertzberg model; the different densities would allow fresh waters to overlie the sea water. Its thickness under the sea level would be equal to 40 times its elevation.

The water table levels are generally modest, 3-5 m and about 1 m north and south of the rift, respectively. The static level increases by about 0.5 m/Km in the northern and by about 0.2 m/Km in the southern area. These differences have been attributed to the barrier effect of the dikes (Druecker, Fan, 1976).

The distribution of permeability at depth in the rift is not well known. Indeed, in the rift it is conditioned by the presence of tensional structures such as small graben, faults, fissures and dikes. The latter, thin and nearly vertical, constitute one of the main factors of discrepancy with the Ghyben-Hertzberg model. Indeed, the dikes isolate blocks of rocks and act as a barrier against transverse flow, locally allowing the water level to rise. According to Druecker, Fan (1976), the phenomena of impoundment between the dikes occurs only in the Upper and Middle Rift.

In addition, the occasional presence of impermeable layers (mainly ash) between the lava flows results in the formation of perched aquifers. The springs and wells that drain these aquifers are located mainly in the Hilo District, in an area where the water levels are very high (250-500 m a.s.l. - Kurtistown, Mountain View 1 and 2), in total disequilibrium with the surrounding water levels. However, these aquifers are of limited importance, as shown in tests of water supply wells (Dreucker and Fan, 1976).

Another factor of discrepancy in the high temperature zones of the rift is the warming up of the deep sea water.

The consequent decrease in density complicates the actual situation with respect to the Ghyben-Hertzberg lens model (Furumoto and others, 1977).

Figure 2/P2 shows the probable distribution of the different kind of waters [perched waters, dike impoundments, basal waters], according to both the Ghyben-Hertzberg model and the particular geological-structural situation of the rift (Druecker, Fan, 1976). The main flow of water in the rift is vertical and in west-east horizontal direction.

The base aquifer discharges along the coast by springs and diffuse flow. Towards the SE the flow of the water coming from the rift is constrained by the presence of dikes.

The residence times of the shallow waters have been estimated at only a few decades on the basis of tritium and $\delta^{18}O$ (Kroopnick and others, 1978).

5.2.4. NUMERICAL MODELLING OF SHALLOW WATERS

On the basis of the known hydrologic characteristics (rainfall, infiltration as estimated from some well data, etc.) Imada (1984) has elaborated a numerical model of the underground flow of waters in the Puna area.

By assigning smaller permeability values to the rift zone than to the regional values of the external zones, it is possible to obtain a good calibration of the model. In particular, the best results are obtained when different permeability values are assigned to the western, central and eastern areas of the rift ($9{,}8 \cdot 10^{-8}$ cm^2, $9{,}8 \cdot 10^{-7}$ cm^2, $9{,}8 \cdot 10^{-6}$ cm^2) (Fig. 3/P2). The literature considers that secondary permeability, due to fracturing, is influenced by the particular structural situation of the rift. The buttressing effect of Mauna Loa on Kilauea is much stronger in the upper zone than in the lower zone. In fact, the Upper East Rift must expand mainly in a southward direction, while further to the east it could expand symmetrically without any constraints. This phenomenon and the presence of a smaller number of dikes in the eastern end of the rift would allow a higher permeability. This higher value would be almost equal to that of the areas outside the rift (4,5 . 10^{-5} cm^2).

The model has been calibrated on the basis of data from the existing wells, which are far too few (Fig. 3/P2). The accuracy of the model is limited by a number of arbitrary assumptions (permeability and dispersivity, heads, isotropic medium, existence of the Ghyben-Hertzberg lens over the entire area, etc.). For this reason, it is necessary to be very careful in evaluating results (Imada, 1984).

5.2.5. TEMPERATURE DISTRIBUTION OF SHALLOW WATERS

The thermal data regarding the shallow wells are very dishomogeneous. Only in a few wells continuous thermal logs have been performed. It is not clear whether the thermal measurements are stabilized or if they were carried out in casing or in open hole. The data collected are not suitable to be processed so as to reconstruct significant geothermal gradient or heat flow attitudes.

The main thermal measurements are shown in Table 1/P2. Figure 4/P2 shows the most significant thermal profiles of a few wells. The maximum temperature in the GTW2 well is 96°C measured approximately at 205 m a.s.l., in the absence of the water level and in evident presence of steam. In the GTW3, a thermal inversion is evident, with approximately 93°C at the sea level and about 62°C at -30 m b.s.l. The temperature measured near the sea level for MALAMA KI was about 55°C, with a modest underlaying thermal inversion.

The reconstruction of the temperature attitude at sea level is an attempt to homogenize the data (Plate 2). For this reconstruction, data from the deep wells have also been used.

In the large area north of the KERZ, the water temperatures (15-22°C) measured in water supply wells are slightly lower than the local average air temperature (=22°C). The area of the rift presents the highest temperatures (about 35°C in the easternmost part, about 50°-65°C in the area of the HGP-A well and nearly 100°C in the GTW2 and GTW3 wells).

The highest temperatures are present in isolated spots which cannot be directly related to each other. This might be due to the more or less punctual upwelling of warm fluids which would be responsible also for the local thermal inversions (GTW3, MALAMA KI).

The thermal inversion of the GTW3 well has been correlated to the density ratio inversion between the overlaying worm sea water and the cooler meteoric waters (Buddemeier and others, 1976; Thomas and others, 1979).

The area south of the rift has temperatures higher than those of the area north of the rift, with values ranging from 26° to 50°C. This is probably an effect of modest lateral leakages of warm water from the rift.

5.3. GEOPHYSICS

The geophysical data available refer to a great number of geophysical methodologies which can be separated into three principal categories.

First are all the general studies on a regional scale with tectonic structural reconnaissance goals. They are constituted mainly by gravimetric and magnetometric data.

A second category is constituted by data referring to natural seismicity and which since the early 1960's have been continuously gathered. These data are important for regional tectonic interpretations as well as for their more local geothermal implications.

The last category comprises all the numerous studies which have been performed for purposes of specific geothermal exploration. These studies deal with electric and electromagnetic methods and with some local active and passive seismic surveys.

Since the data was obtained from reports or otherwise printed literature, they are not in the form in which they were originally collected. Given the technical and scientific status of the authors we will consider that the elaboration and interpretation of the data are the most advanced and reliable. For these reasons, rather than a new interpretation, we will only provide a synthetic and comparative analysis of the most significant data available.

5.3.1. GRAVITY DATA

A "Bouguer Anomaly Contour Map" (Kinoshita, 1965) of the entire Island of Hawaii collects the data of the several surveys performed by the U.S.G.S. during the years 1961-1962. This map is the most significant gravimetric study on a regional scale.

Access to vast areas of the island is hindered by topography and by the dense vegetation. The few gravimetric stations, on whose data the map was elaborated, are therefore not distributed homogeneously but only along the existing highway network.

Recently (Furumoto and others, 1976) in the context of a geothermal project, a detailed gravimetric study has been carried out in the KERZ. This study consisted of 210 new stations distributed in an area of approximately 150 km^2.

Both the regional survey map of the Island of Hawaii as it refers to the KERZ and the more detailed map of the Puna District are respectively shown in Plate 3.

The gravity values of the detail map (Fig.1/P3) are very different from those reported by the regional survey, because, in elaborating this map, the gravity values were reduced. The maximum value assigned in the area was "zero". The form of the anomalies, instead, is similar except for small variations which are due to the greater detail of the survey.

In particular, the presence of a wide positive anomaly oriented along the axis of the KERZ is confirmed. The map of the regional survey indicates that the maximum value of this anomaly corresponds to the Kilauea Crater.

On the basis of the data provided by the detail survey, quantitative interpretation based on the bidimensional modelling (Broyles, 1977) of the two profiles shown in Fig.1/P3 has been made. The same figure shows the matching between the observed and calculated data for bodies with different density contrast and form (Fig.2/P3).

The fact that the fitting does not show considerable variations underlines the great degree of ambiguity inherent in the interpretation. However, for both profiles the best fitting is obtained for bodies with a density contrast between 0.4 - 0.5 g/cm^3 and depth of the top ranging from 2 to 3 km approximately. Only along the B-B' profile, and for a limited area corresponding to the maximum of the anomaly, does such top reach near the surface.

Since the gravimetric interpretation must satisfy not only the observed gravimetric field but also the hypothetical geological-structural model, it is possible to associate the body causing the anomaly to the set of intrusions that constitute the Dike Complex, whose composition corresponds to the effusive basaltic-olivine products.

The direct measurement of the densities (Strange and others, 1965) of some olivine basalts collected throughout the Island of Hawaii has produced values ranging from 2.8 and 3.1 g/cm^3.

A density value of 3.1 corresponds to the density deduced from the velocity/density relation (Manghnani and Woollard, 1968) obtained from seismic studies. Indeed a layer with a seismic velocity of 7.0 m/sec 10^3 has been identified and may correspond to a deep formation of intrusive type. A density of 2.5 - 2.7 g/cm^3 for a shallower and slower (5.1 - 5.7 m/sec 10^3) layer has been deduced from the same seismic surveys.

The density contrast of 0.4 - 0.6 for the gravimetric modelling was estimated on this basis. This is similar to what has been shown for the similar rifts of the Koolau lava shield Volcano of Oahu (Walker, 1986).

The contribution of the available gravimetric data, hence, is that of having supplied useful indications about the deep lateral extension of the Dike Complex (12-20 km), identified as the body causing the positive gravimetric anomaly of the KERZ.

However, the gravimetric interpretation is limited by the extreme lack of homogeneity in the distribution of the measuring stations, by the small number of direct density measurements and by the value of surface density (2.3 g/cm^3) used for the Bouguer correction. This value may be too low, as a correspondence between the topographic setting and the contouring of the anomalies shows (Ross, 1982).

Nevertheless, from a qualitative point of view the information furnished by the detailed survey of the Puna District area is important. As can be seen in the corresponding map (Fig.1/P3) the positive gravity anomaly shows a northwards displacement of its easternmost part between Pahoa and Cape Kumukahi.

Unfortunately if analogous situations exist in other sectors of the KERZ, it is impossible to detect them because, as already mentioned, detailed data are either not homogeneous or do not exist.

5.3.2. MAGNETIC DATA

The main magnetic data available refer to two aeromagnetic surveys made in 1966 and 1978.

The 1966 survey covered the entire Hawaiian archipelago. To better understand the submarine volcanic environment, it was integrated (Malahoff and Woollard, 1968) with a shipborne survey over the entire underwater extension of the Kilauea East Rift. The magnetic map plotted with these data was made through the collaboration of the USGS.

Because of the regionality and the high flight altitude of the survey (approximately 4000 m), the results are significant only for the identification of very large and deep structures. For this reason the quantitative interpretations elaborated upon these data (Furumoto, 1978) are of limited interest to characterize shallower structures, such as Dike Complex.

The 1978 survey, carried out by the USGS, covered only the Island of Hawaii. The data were obtained in flights at an altitude of 300 m and, in the Kilauea area, along lines oriented N 30°W and spaced 0.8 - 1.6 km. This gives the survey a degree of detail even where structures close to the surface are concerned. Among the available data this survey is the most significant and up-to date. In particular, the first published map of the surveyed magnetic anomalies (Godson and others, 1981) was made subtracting the International Geomagnetic Reference Field (IGRF) from the observed field.

More recently (Flanigan and others, 1986), a detailed map of the area of the rift systems of the Kilauea and Mauna Loa volcanos has been elaborated. The map was based upon data gathered in 1978 by means of low altitude (100 m) magnetic and electromagnetic surveys (Very Low Frequency). This map shows the total magnetic anomalies, without subtracting the IGRF, because at the small magnetic latitudes of the Island of Hawaii the regional field does not exceed 4 nT/km. Hence the anomalies do not change significantly, either in their form or in their intensity. Plate 4 shows the portion of this map that refers to the Kilauea East Rift. The most significant elements are the net linear magnetic gradients and the important dipolar magnetic anomalies which, aligned along an approximately E-W trend, closely follow the series of cones, craters and fissures of the KERZ.

Considering the dip of the magnetic field (35° N) at the local latitudes, the orientation of the dipolar anomalies is in accordance with a direction of the remanent magnetization of the source bodies close to that of the present magnetic field. Therefore the emplacement, subsequent cooling down and magnetization of these bodies occurred after the last magnetic inversion ($2*10^4$ y) and hence according to the present magnetic field of the Earth. All these considerations allow to identify the source body with the basaltic intrusions constituting the Dike Complex and which, with an E-W trend, grow the southern flank of the Kilauea East Rift.

The quantitative interpretations of several dipolar anomalies (Ross, 1982; Flanigan and others, 1986) have furnished indications as to the form and depth of the causative bodies. In particular, the best fitting has been obtained for sub-surface bodies (top to maximum depth a few hundreds of meters from the g.l.), with a length ranging from 1-2 to about 10 km and with a magnetic susceptibility contrast of about $1*10^{-1}$ cgs units (Fig.1/P4).

In addition, the lateral extension of these source bodies seems to coincide with the maximum and minimum values of the dipolar magnetic anomalies. Therefore, as Plate 4 shows, the borders of the source bodies can be traced. This is true especially where the magnetic anomalies are well marked and evident, as in the KERZ.

However, the reliability of the quantitative interpretation can be compromised by the insufficient knowledge of the magnetic susceptibility values characterizing both the causative intrusive bodies and the host rocks. Indeed, measurements of magnetic susceptibility and of natural remanent magnetization were taken only on 30 rock samples gathered from the entire Island of Hawaii (Malahoff and Woollard, 1965).

Both for the tholeiitic and for the recent basaltic-alkaline samples from Puna District, the susceptibility values are of the same order of magnitude ($1.54*10^{-3}$ cgs units and $3.62*10^{-3}$ cgs units respectively).

However, it has been observed that the samples belonging to the same lava flows can present variations in their magnetic susceptibility of up to $1.0*10^{-1}$ cgs units, depending on the absence or presence of local concentrations of ferromagnetic minerals.

Undoubtedly these data provide a regional view of the most representative susceptibilities but they cannot be representative for the KERZ. This is so especially because there are no direct measurements from dike samples from the KERZ.

Taking into consideration the available modellings and interpretations it is still possible to highlight other significant aspects of the map reproduced in Plate 4.

Several circular anomalies are evident, in addition to the already discussed dipolar magnetic anomalies which are normally polarized, elongated in the direction of the KERZ, and associated to the Dike Complex. Some of them have an inverse polarization. This can suggest that the causative intrusive bodies were emplaced in periods in which the direction of the magnetic field of the Earth was opposite to its present direction. This, however, disagrees (Malahoff and Strange, 1965) with the age of the volcanic apparatus of Kilauea, the shallow products of which would be younger than the most recent inversion of the magnetic field of the Earth.

Theoretical calculations have shown that, at least for some of these circular anomalies, the apparent inverse polarity is the result of topographic factors. This is the case of the anomalies associated with the Halemaumau and Kilauea Iki craters.

The case of the Kapoho crater is different. Here the presence at a shallow depth of a non magnetic structure, which probably may be associated to a magmatic camera with temperatures above 580°C (Curie point), has been hypothesized. The corresponding model (Flanigan and Long. 1987) is shown in Fig.1/P4.

The map also highlights the interruption of the E-W oriented linear magnetic anomalies of the KERZ north of Opihikao. This element, which has been interpreted as a tectonic structure transverse to the direction of the rift, corresponds with the northwards displacement of the positive gravimetric anomaly.

5.3.3. SEISMIC DATA

The seismic data available are fundamentally seismological (passive seismic). Few data, in fact, have been obtained from seismic refraction profiles.

The seismological data provide useful information on the volcano-tectonic setting of the Kilauea volcano, one of the most active zones in the world from a seismic point of view.

Seismicity of the Kilauea volcanic complex

A network of seismometers has been in operation on Kilauea since the 1950's. The number of seismometric stations has increased, and the quality of the instruments and of the programs to determine the focal parameters of the earthquakes have improved with time.

By 1983 the configuration of the seismic network managed by the HVO (Fig.1/P5) already permitted a good coverage of the entire KERZ (Klein and others, 1987).

Because of the improvements to the seismic network it has been possible to survey a much greater number of earthquakes and to noticeably improve the hypocentral determination.

In the early 1960's the quakes could be located only if they were of magnitude greater than 2; today the level of detectable magnitude is around 1. This development has increased the number of events that can be recorded about tenfold. The standard error of the focal coordinates determination has improved from ≤ 2.5 km to 0.5-1 km. All of this has made it possible to study quake swarms and to correlate them better with even small magmatic intrusions.

Particularly for the Upper East Rift there is abundant literature which clearly demonstrates the correlation between Kilauea's seismic activity and the change in the stress conditions brought about by volcanism. In contrast, the seismic swarms coming from the lower part of the East Rift cannot be correlated to specific volcanic events (Koyanagi and others, 1981).

The very high number of seismic events recorded and processed since the early Sixties (over 15000 only in the 1970-1973 period) does not allow for a clear graphic representation in a single map. However, the epicentral maps elaborated by the HVO and available through the Bulletin of the Observatory since 1962 show that the maximum concentration of crust earthquakes occurs in the southern slope of Kilauea (Fig.2/P5).

The identification of seismically homogeneous zones based upon the events recorded from 1968 to 1983 (Klein and others, 1987) can provide a characterization of the seismicity of the whole KERZ. The entire area of Kilauea and of the East Rift has been divided into seven principal areas (Plate 5), whose cumulative curves of earthquake occurrence are shown in Fig.3/P5.

The Kilauea caldera area is characterized by a first zone with very high shallow seismicity (6000 events between 0-5 km), a smaller zone with lesser and deeper seismic activity (about 500 events both between 5 to 13 km and 13 to 20 km) and a third, wider zone characterized by a very deep and more intense seismicity (about 750 events with M $\geq$ 2 between 20-60 km). This latter zone is certainly associated with the deep main magmatic chamber. Indeed, the corresponding cumulative curve does not show the step trend typical of more shallow seismicity which is clearly correlated to frequent volcanic effusive or intrusive episodes.

For the East Rift the cumulative curves show how the shallow (0-5 km) seismicity is far more active for the Upper East Rift (about 1000 events) than for the Middle East Rift (about 350 events). The difference in the seismic activity of greater intensity (M $\geq$ 2) is even greater between the South Flank zone (over 7000 events) and the Lower East Rift ($\approx$ 550 events). Finally, it is interesting to note how, after the strong earthquake in 1975 at Kalapana (M=7.2) the medium deep seismicity of Kilauea was strongly diminished, while that of the South Flank and of the Lower East Rift increased considerably.

A study of events with $M \geq 3$ during the 1975-1977 period and of the aftershocks of the Kalapana earthquake mentioned above (Koyanagi and others, 1981) shows (see Fig.4/P5) a heavy concentration of epicenters on the southern flank. Instead, the areas north of the East Rift, as well as those NE of Puu Honuaula, are quite aseismic. Most of the events occur at the depth of 5-10 km near the base of the volcanic pile (Swanson and others, 1976). In contrast, events diminish rapidly beneath 10 km except for the area under the southern summit of the volcano, where earthquakes occur up to a depth of about 60 km.

The asymmetric distribution of the crust earthquakes with respect to the axis of the East Rift is correlated to a release of the stresses generated by the magmatic intrusions. Such release occurs preferentially relative to the free slopes towards the sea (Koyanagi and others, 1972).

This mechanism, which has been proved by detailed surveys of ground deformation (Swanson and others, 1976), occurs because the northern flank of the rift buttressed by the massif Mauna Loa, remains stable, is almost immobile and hence relatively aseismic.

To better illustrate the tectonic setting of a region with natural seismicity data, strain release maps have been made (Fig.5/P5). The contours of these maps (Koyanagi and others, 1981) are strongly influenced by few earthquakes of larger magnitude. During the period under consideration the greater quantity of energy was released along a seismic strip that extends for 60 km parallel to the southern flank of the East Rift at a depth of 5-10 km. At greater depths the entire area appears quite aseismic and therefore stable.

Up to a depth of 5 km, the subsurface seismicity is located south-east of the caldera of Kilauea. In this area earthquake swarms, which may be associated with magmatic intrusions, occur constantly. The summit of Kilauea and the adjacent rift zones are also characterized by swarms of small shallow earthquakes (0-5 km) associated to the volcanic activity.

Microearthquake surveys

Some microearthquake surveys have been adopted in the Lower East Rift as a method for geothermal research.

The surveys were carried out for very short periods of time with seismometer arrays that could register seismic events with a magnitude of up to -1. For most events the error in the location, both horizontal and vertical, was of the order of 1 km. A group of microearthquakes occurred in the first 3 km of depth, exactly in the location of the HGP-A well (Fig.6/P5), was correlated to the hydrothermal activity of the geothermal reservoir in exploitation (Suyenaga and Furumoto, 1978; Mattice and Furumoto, 1978).

Another cluster of epicenters (Plate 5), interpreted as a marker of geothermal reservoir, was evidenced between Kalapana and Pahoa (Mattice and Furumoto, 1978). The reliability of these data is limited by the short period of recording (20 days) and by the surveying pattern adopted (only 2 seismometers employed).

A seismic ground noise survey of the same area was also carried out. It showed that the ground noise anomalies are not spatially correlated to the geothermal activity but can be related to the volcanic features of the KERZ (Norris and Furumoto, 1978).

However, especially in a geodynamic context as complex as that of the KERZ, on the basis of these studies it is nearly impossible to distinguish the seismic activity generated by hydrothermalism from that caused by the extremely diffuse magmatic activity.

Seismic refraction surveys

As to active seismic surveys, some seismic refraction surveys with shallow investigation depth were made in order to obtain a reconstruction of the shallower structures of the Lower East Rift (Suyenaga, 1978).

A later survey, with much longer refraction lines, allowed to determine a body with a 7.0 m/sec*10^3 velocity associated to the Dike Complex. Its top is located at a depth of 2-2.5 km (Fig.7/P5) while its bottom has been located by other studies at about 5 km. Furthermore, depthwise the lateral extension of the rift, the dimensions of which range between 12 and 19 km (Broyles and Furumoto, 1978), is far wider than its surface expression. This is consistent with what has been demonstrated by gravimetric data.

This geologic-structural model of the Lower East Rift, reconstructed along a N-S profile comprising the geothermal reservoir identified on the HGP-A well, has been made using the data furnished by the seismic refraction profiles, ground noise and microearthquake surveys already cited as well as other geophysical data available (Furumoto, 1978).

5.3.4. GEOELECTRICAL DATA

For the Kilauea East Rift there is a vast documentation of geoelectrical data generated by numerous studies performed as part of geothermal research and with many and diverse exploration techniques. Electric and electromagnetic methods are the most used.

It is wellknown that the resistivity of a terrain diminishes as its porosity, salinity and the temperature of its circulating fluids increase, as well as with the concentration of minerals associated to hydrothermal alteration. Therefore, such studies are oriented towards the identification of electric conductivity or resistivity anomalies.

The wide range of techniques adopted is due to the need to resolve or minimize the main operative problems encountered in the area (high electrode impedance, presence of numerous pipelines and electroducts, which alter the values observed, total dispersion of the current consequently to the high conductivity due to sea water "intrusions".

Dipole-dipole techniques

The first of this kind of surveys in the area consisted of 13 Dipole-Dipole soundings (Keller and others, 1977) distributed over the entire area of the Kilauea Crater and its East Rift. These soundings (Plate 6) used a current Dipole 1-3 km long.

For logistic reasons, the distance between source dipole and measurements dipole was not gradually and homogeneously increased in order to have a major depth of exploration. The correlation of data supplied by different soundings was thus problematic and ambiguous; their interpretation could only be general and cannot generate clear indications on the vertical or horizontal variations of resistivity. Therefore, it was possible to reconstruct only qualitative "resistivity-thicknesses" models for the area studied by any D-D sounding (Plate 7).

A second study (Keller and others, 1977) concentrated on only the Lower East Rift. It used a pole-dipole device to determine the variations of apparent resistivity as a function of depth along three profiles (Plate 6). The depth of investigation of the three sections thus constituted was limited to about 600 m. In general, the sections showed that the top of a low resistivity stratum (≤5 ohm*m) could be identified at depths greater than 500 m only in the area of the Kapoho Crater. Even in this case the data provide only limited qualitative information. For this reason both the resistivity and the depth data obtained should be considered as very approximate and with little degree of correlation.

Electromagnetic Soundings

Several electromagnetic time domain soundings were performed in the KERZ (Keller and others, 1977 - Skokan, 1974) to complete the soundings made with the Dipole-Dipole reconnaissance survey. The location of the source lines is shown in Plate 6. In Plate 7 and in Fig.1/P 7 the traces and the corresponding geoelectric sections are respectively shown. The latter were reconstructed with the data mentioned previously. There is a general agreement between the two studies, but detailed data are lacking.

Twenty-six EM soundings at an effective penetration depth of ≤ 200 m were made in the area of Puna District. These soundings determined only the depth of the interface fresh water/seawater, thus making a modest contribution to the geoelectric characterization of the area.

An important set of data in the LERZ is an EM Transient survey (Kauahikaua and Klein, 1977), consisting of twenty-four soundings made in an area of 50 km^2. Only seventeen of these (Plate 6) were interpreted with a layered model. The data processing did not give very reliable contourings of the resistivity and thickness of the conductive layer underlying the surface resistive cover. Nevertheless, both contourings are shown in Plate 7 and it is possible to note that the resistivity low (2 ohm*m) of the conductive layer includes the HGP-A well and the area of the Puu Honuaula Crater.

Low resistivity values (4 ohm*m) at greater depths are explained by the temperature factor estimated as being of 200-250°C. These soundings provide rather reliable data only for a small area of the Lower East Rift and only for investigation depths not below 1000 m.b.s.l.

A detail survey with greater research depth (Kauahikaua and Mattice, 1981) included 6 TDEM sounding (Time Domain Electromagnetic) south of Pahoa and 5 in the area of Keaau. These soundings (Plate 6) have confirmed the data advanced by the preceding 1977 study on the surface conductive layer by highlighting the existence of an underlying resistant basement. Each of the apparent resistivity curves thus reconstructed (Fig.2/P7) shows the top of the deep resistant structure to be under 3 km.

VES (Vertical Electric Sounding)

One of the first electrical surveys was carried out on the Island of Hawaii during the mid-1960's using the VES method with Schlumberger configuration and equatorial direct current dipole (Zohdy and others, 1965). The goal was to determine whether direct current methods can be used in cold aquifer research. The project demonstrated the applicability of Schlumberger electric soundings in conditions of high surface resistivity even with the 4-6 km AB device (AB = electrode spacing).

In 1979 a group of USGS and HIG geophysicists (Kauahikaua and Mattice, 1981) surveyed an area of well-known geothermal interest. The survey consisted of 15 VES (Plate 6), the 13 TDEM soundings mentioned above, and a self potential profile. The maximum current electrode spacing was 2000 m. The investigation depth, therefore, was not greater than 500-700 m.

The surface resistivity contrast between water saturated rocks and dry rocks was again demonstrated by this study. In particular, it showed that the rocks lying above the water table typically have resistivities of the order of thousands of ohm*m while the rocks under the water table have resistivities that may even be below 5 ohm*m.

Apart from the fact that this study had shallow penetration, the resolution of the method is conditioned by the ocean effect. In fact, rather than penetrating into the underlying rock layers, the electric current introduced into the ground flows laterally in this highly conductive medium.

On the basis of the above electric and

electromagnetic data, a possible differentiation of the Lower East Rift into two areas with different geothermal interest has been suggested (Fig.4/P7). In particular, the subsurface geoelectric differences between those areas are due to hydrogeological differences and Area II would be characterized by a surface fresh water lens thicker than that found in Area I. The latter is indicated as the most promising area for geothermal exploitation (Kauahikaua and Mattice, 1981).

The deeper geo-electric structure of the LERZ is attributed, mostly, to the effect of heat on the water saturated rocks. The abnormally conductive layer has been interpreted as water bearing with temperatures above 200°C. The bottom of this layer has been mapped to a depth of 1000-1300 m b.s.l. on the flanks of the rift and to 250-500 m b.s.l. the rift and near the HGP-A well (see Plate 7). This data is not confirmed by a thermal profile performed in the HGP-A well.

The resistive basement has been interpreted as the result of a reduction of the porosity linked to the pressure conditions at a depth of 3 km.

Mise à la masse

A mise-à-la-masse survey was carried out in the area of the HGP-A well using the casing of the well and an electrode positioned in the sea as current electrodes (Kauahikaua and others, 1980). Along the existing roads the electric potential up to a distance of 2 km was measured.

The electric potential measurements were transformed into apparent resistivity values (Fig.5/P7). The data processing showed the existence of a zone with resistivities ranging from 10-20 ohm*m in proximity of the HGP-A well and of a zone with lower resistivity values (2-5 ohm*m) encircling it. The zone with higher resistivity was explained as due to a group of dikes fresh water impounded and surrounded by formations in which fluids with high salinity and temperature values circulate.

Even if the ambiguity of this interpretation is considered, it is noted that all the drilled productive wells are located inside the electric potential anomaly.

Self potential survey

Two self potential surveys were carried out (Plate 7). The first was carried out at the Kilauea volcano and extended to the margin of the Upper East rift (Zablocki, 1976). The second was made in the Lower East Rift (Zablocki, 1977).

As it is known, natural electric currents may originate through the differential movement of ions in thermal waters (electrokinetic mechanism) or through thermoelectric effects present in geothermal areas.

The potential anomaly that has been found at Kilauea is related to the local heat sources, while anomalies in the East Rift, including the HGP-A geothermal well, are due to convective movements of the geothermal fluid in the area. These anomalies, however, are linked to shallow causes.

Airborne Electromagnetic Survey (VLF)

The only resistivity study that has homogeneously covered the entire area of the Kilauea Crater and its East Rift is an Airborne Electromagnetic Survey with VLF (Very Low Frequency) modality (frequency range 18-24 kHz). This study (Flanigan and others, 1986), was made by the U.S.G.S. in 1978 as part of the aeromagnetic survey cited above, and allowed to prepare an apparent resistivity map (Plate 8) for a frequency value of 18.6 kHz.

Therefore, considering that the surface resistivity can vary in this area from 100 to 10000 ohm*m approximately, the skin depth of an electromagnetic wave varies, in turn, between 36 and 360 m. Hence the resistivity anomalies are linked to very shallow causes and structures.

In general in the East Rift the shallow resistivity values range from 60 to 1600 ohm*m. The lower values identified along the northern flank and the southern coast can be attributed, respectively, to the interference of electroducts and to the great difference in the resistivity values of rocks and sea.

It is possible, however, to identify some low resistivity areas in the KERZ which are important.

The most significant one among these areas extends from the coast towards Puulena Crater with a NW trend. Here a Self Potential anomaly (Zablocki, 1977) attributed to a sub-surface hydrothermal convective cell is present and could be connected to a deeper hot source; this might be attested by a small magnetic low centred on Puulena Crater (Plate 4).

In contrast, a specific low resistivity does not correspond to the apparently reversed magnetic anomaly of Kapoho Crater. This may be due to the recent volcanic activity of Kapoho, which would have masked all the shallow resistivity anomalies associated to the presumed magmatic chamber (Flanigan and others, 1986). Nevertheless, the resistivity data from the Dipole survey (Skokan, 1974) suggest the probable existence of a deep conductive anomaly in the area of the Kapoho Crater.

The southeasterly propagation of the resistivity low of Puulena is associated to a lateral descending circulation of hydrothermal water of meteoric origin. This would be confirmed by the warm springs that are found along the coast and by the hot water (55°C at a depth of 90 m) found in the Malama-Ki well.

This anomaly could be interpreted, as the other geophysical elements previously mentioned, as a transverse structure of the rift which, by breaking the impermeable dikes, controls the water circulation.

East of Napau Crater and west of Kalapana there are two similar conductive anomalies transverse to the axis of the rift. They may represent two transverse fractures caused by the intense differential block movements of the southern flank of the East Rift. All these fractures associated to the conductive anomalies would constitute preferential paths for sea water inflow. However, clear correlations with other structural surface elements or with other geophysical indicators are lacking.

5.4. GEOCHEMISTRY

On the basis of the location of the KERZ, we have considered data on water points located between 154°47'30" and 155°30' Lat. W and between 19°07'30" and 19°45' Long. N. This area should be sufficiently large to allow an exhaustive outline of characteristics of the fluids and to define tentative hypothesis about deep circulation.

5.4.1. AVAILABLE DATA

The data available in the literature refer predominantly to two kinds of samples:

- fluids from shallow wells (< 300 m deep) for thermal measurements and water supply;
- fluids from deep geothermal wells.

The data from reports and numerous publications are frequently fragmentary and incomplete. In particular:

a) The chemical analyses are limited to the main elements; trace elements are determined only in a few cases. The latter are necessary for a complete classification of the waters and for a good outline of the geothermal fluids deep circulation.

b) Stable isotopes determinations are very few and random: information on the content of 18 oxigen and deuterium are essential to define the relationship between the geothermal component and fresh water. In this area further difficulties arise for interactions of both deep and shallow aquifers with sea water.

c) The situation is not better concerning the HGP-A well, although it has been producing for many years. Indeed, there are no complete sets of samples (separated liquid, condensed steam and gas collected contemporaneously) that allows the characterization of the fluid in reservoir conditions. While there are numerous analyses of separated liquid and gas there is a total lack of condensate analyses. Furthermore, if the first production tests are excluded, measurements of flow rates and enthalpy are also totally lacking. No systematic isotopic analyses are available except for in-hole samples of dubious reliability collected before the recasing operations in November 1979.

5.4.2. QUALITY OF THE DATA

After having pointed out the gaps in the data, an evaluation of the quality of the available ones is required. The chemical and isotopic analyses of shallow water points and deep wells, except for HGP-A, are shown in Tab.1/An.A. The HGP-A data are in Tab.2/An.A. Of the 94 samples given in Tab.1/An.A, for which the principal elements were determined, only 47 have unbalances between cations and anions below 10%.

There is uncertainty about the location of some water points; in the documentation examined, their coordinates are not defined unequivocally. Some of them present variations, sometimes considerable, regarding the temperature, chemistry, and salinity. It is not clear whether these variations are due to real changes occurring with time or to the sampling in different points (for example, several wells close to one another and with the same name).

It is not known if the alkalinity values refer to field or laboratory measurements. In the latter case, the bicarbonate concentration might be slightly altered and hence contribute to the analytical unbalances.

5.4.3. CLASSIFICATION OF THE SAMPLES

The chemical characteristics of both shallow waters and deep well samples are examined here below. The water points location are shown in Fig.1/An.A. In the following figures of the same Annex, the samples are identified with the corresponding codes used in Tab.1/An.A.

Samples with low chloride content (< 35 mg/l).

Of the 41 samples belonging to this group, only 12 have an unbalance under 10%. By increasing the threshold to 15%, the number of significant samples reach 25 (Tab. 3/An.A). Only the latter, which are distributed mainly in the area of Keaau, Puna and Kau have been considered in this report.

Their temperature ranges from 17.8 and 28°C, the T.D.S. is around 100-150 mg/l and the SiO_2 content varies between 21 and 55 mg/l. Their chemistry is mainly alkaline-bicarbonate alkaline-earth bicarbonate; the contribution of these components is greater than 60%. The alkaline-earth sulphate fraction varies between 15 and 25% and the alkaline-chloride component ranges between 5 and 25%. Only the Kanoelehua 2 (10a), the Kanoelehua 3 (11a) located in Hilo, and the Hawn Shores 2 (2, 3), 5 km southeast from Hilo, have the alkaline chloride component around 50%. In the square Piper diagram in Fig. 2/An.A, all the samples with a high alkaline-earth bicarbonate component are located in the northwest quadrant. The other samples, in which the contribution of other components is higher, are shifted. Their "mixed" character is confirmed by the modified Piper diagram of Fig. 3/An.A. Here the samples Pahoa 1 and 2 (33, 34 and 34a) are clearly located in the southwest quadrant, while slightly to the right there are the samples Hawn Shores 1 and 2 (2, 3). The composition of these very diluted water points, all located to the north of the deep wells, might be explained by the presence of gas effluents (CO2 principally) that permeate the aquifer nappe.

In conclusion, according to the chemistry, salinity and temperature, all these samples can be considered as rather homogeneous fresh waters occasionally characterized by a gas inflow.

As regards the three samples where the alkaline-chloride component is high, even if located near the coast, the examination of the Cl/Mg vs. Cl diagram (Fig.4/An.A) excludes any sea contamination. In fact, all the water points show a low Cl/Mg ratio and are extremely well correlated, according to a common surface leakage process (marine spread).

<u>Samples with high chloride content (> 35 mg/l</u>)

Of the 53 samples with chlorides above 35 mg/l, 35 have an analytical unbalance < 10%, and 41 < 15%. The Allison sample (1b) is not henceforth considered because normally five to eight times more concentrated.

This resulting set of 40 samples can be, for practical and conceptual purposes, divided into two sub-groups:

- sub-group (A), T.D.S. < 1150 mg/l, Cl < 600 mg/l
- sub-group (B), T.D.S. > 2000 mg/l, Cl > 1000 mg/l.

<u>Sub-group</u> (A) consists of 22 samples collected in 15 different locations. Their temperatures vary between 18 and 28.5°C excluding Kapoho LSW (13, 13c and 13d) for which a temperature of about 37°C has been measured. In general, the SiO_2 content is close to that of the samples previously discussed; only the Kapoho LSW and the Pulama samples show a noticeable increase up to 70 mg/l. The chemistry is in general alkaline-chloride; this component varies, in fact, between 57 and 78%; only the Kapoho CS samples (12 and 12a) are alkaline-bicarbonate.

In the Piper diagram (Fig. 5/An.A) all the representative points, with the exception of the Kapoho CS ones, are located in the southeast quadrant characteristic of alkaline-chloride waters.

As regards the Cl/Mg vs. Cl diagram (Fig.6/An.A), the samples containing more than 3 meq/l of chloride, except for Keauhoana 1 (21, 21a, 21c and 21d) and Kapoho CS (12a), have been certainly contaminated by seawater. Their Cl/Mg ratio rises quickly to the seawater one as the chloride content increases. The original composition is no longer recognizable but it should be connected with the "fresh waters".

The Keauhoana 1 is related to the same leaching process that characterizes the fresh waters being located on the same regression line. The Kapoho CS representative points are very far from the other samples with a comparable chloride content. This peculiarity together with the high alkalinity value make them a singled out set with a different composition even if they are located near Kapoho LSW and GTW-4.

Finally, the other samples Waiakea 4 (45) and Keaau Orch (19, 19a and 20a) can be considered as originally earth-alkaline waters with a seawater contamination of few per mil.

In summary, none of the sub-group A samples is related to a deep geothermal circulation.

Sub-group B consists of eighteen samples. Their chemical composition together with a seawater analysis for comparison is given in Tab.5/An.A.

Five come from 3 deep wells: HGP-A, LANIPUNA 6 and KS-1A, where measured temperatures range between 170°C and 360°C; 5 others were taken from the test hole GTW-3 where the maximum temperature is 93°C. The other samples, except for Isaac Hale spring, are shallow wells drilled for water supply.

Waiakea (44), near Hilo, and Honuapo Mill (8), in the area of Keau, have a temperature of about 19°C; all the others have a temperature constantly above 37°C and are located close the southern boundary of the rift. The T.D.S.

of the group varies between 2300 and 35750 mg/l. The chemistry is highly alkaline-chloride with percentages ranging between 77 and 93%. The SiO_2 content, evidently very high in the deep wells samples, is considerable and correlated with temperature for the GTW-3 (G3, G3e, G3f, G3g, G3h), for the Isaac Hale spring (23) and for the Malama Ki (25) water points. The Honuapo Mill (8) shows a SiO_2 concentration similar to the values measured in the sub-group (A) water points located in the Keau area. On the other hand, the SiO_2 concentration in Allison (1) is lower than the expected value, associated with its temperature.

All the samples of this sub-group are located in the southeast quadrant of the Piper diagram (Fig. 7/An.A). Regarding the Cl/Mg ratio vs. Cl (Fig. 8/An.A), all the water supply wells and the Isaac Hale spring present a ratio of approximately 5 and are certainly related to seawater.

As regards the deep geothermal wells, the ratios are about 350 for LANIPUNA 6, 3600 for KS-1A and 11800 for HGP-A. The GTW-3 samples have intermediate values, about 20. Therefore, for the above samples there is an evident relation between measured temperatures and Cl/Mg ratio. In fact, when equilibrium occurs between host rocks and circulating fluids, the higher is the temperature the lower is the magnesium content.

In conclusion, among the shallow water points only the GTW-3 samples presents a surely deep geothermal inflow; but the lack of chemical trace elements and stable isotopes analyses does not allow to quantify the possible seawater contamination.

The other water points with temperatures which exceed seasonal mean value: Isaac Hale (35°C), Allison (38°C) and Malama Ki (55°C) are to be found in a relatively small area (between 2-5 km from the productive wells). The samples (25 c) and (25 d), which are analytically more reliable for Malama Ki, have a very high sulphate content (around 470 and 600 mg/l respectively). Such a high presence of this component, which is almost completely absent in the deep fluid, seems to indicate contamination varying between 21 and 27% of seawater in this water point. The sodium, potassium, magnesium and chloride contents are, in both samples, congruent with such contributions.

The temperature of 55°C may only be justified by conductive or convective heating. The former could be consistent, considering the rise of nearly 30°C compared to the seasonal mean temperature, only with the presence of a localized and intense source of heat (dike). If more diffused heating of the aquifer is hypothesized, a convective mechanism must be taken into account. In terms of thermic balance, a contribution of 10% of deep fluid with a temperature of 350°C would be sufficient to justify the temperature measured.

On the basis of the available data, provided that the composition and the concentration of this deep fluid lie between the range of values that have characterized the production of the HGP-A well, it would be practically impossible to point out this inflow in the shallow aquifer, highly contaminated by seawater. Similar considerations are also valid for the Allison and Isaac Hale water points where the seawater contamination is similar but the temperature is lower.

5.4.4. PRODUCTION HISTORY OF THE HGP-A WELL

After examining from a general point of view the chemistry of the fluids in the area under study, the first 8 years' production of the HGP-A well (out of a total of 14) will now be considered in greater detail. The chemistry variation with time will have to be studied considering only the major cations and chlorides because alkalinity and sulphates determinations are frequently lacking (Tab.2/An.A).

Chronology of the tests

The HGP-A, which is about 1970 m deep and has been cased to 675 m, was opened to the atmosphere for the first time in July, 1976. The maximum temperature value (358°C) was registered at well bottom; a secondary maximum of 320°C was found at a depth of 1370 m (Kroopnick and others, 1978). The well was recased on November 1979 with a 7 inches casing up to 915 m, excluding in this way the productive level at 680 m. Since then it has not been possible to take any downhole sample. There are, however, partial isotopic and chemical analyses of down hole samples collected in dynamic and static conditions between August 1976 and February 1979.

Many well tests were performed before the power station began operating. They are summarized here below:

1 - July 1976: flashing tests for 4 hours. On August 19th, downhole samples were taken at 305, 1310, 1768 and 1920 m and were analyzed only for 18 oxigen content. These samples should be a mixture of deep fluid and drilling water still present in the reservoir.

2 - November 1976: a two week flow test started on November 3rd with an on-line cyclone separator. Gas, condensate steam and separated liquid analyses are available. The condensate samples, because of their high sodium and silica content, are polluted with entrained liquid. It has been impossible to recalculate precisely the composition of the fluid because the sampling pressures were not reported.

3 - December 1976: a six day flow test was performed between the 12th and the 19th of the month. The samples were taken in the same way as during the previous flow test. On December 2nd and 3rd, 1976, downhole samples at 690, 1310 and 1770 m were taken. On January 25th, 1977, samplings were carried out at 692, 1067, 1676 and 1768 m.

4 - Flow test from January 26th to February 11th, 1977: two isotopic analyses of the top steam were made, but none of the corresponding separated liquid. Downhole well samplings at 690, 1070, 1770 and 1920 m were performed

on February 14th, 1977.

5 - March 28, 1977 to May 9, 1977 flow test: notwithstanding the test lasted for 42 days, only 3 chemical analyses are available. There are flow rate and enthalpy measurements of the fluid produced 25 hours after the beginning of the four last tests. No other measurements of this kind are available.

6 - July, 1977 flow test: isotopic and chemical analyses of a single gas sample were performed. There is no further information available.

7 - June, 1978 flow test: only one chemical analysis of separated liquid is available. There is no further information.

8 - January 3rd to 18th, 1980 flow test: there are chemical and isotopic analyses (tritium) (Thomas, 1980). The isotopic analyses do not seem to be significant.

9 - June 11th to September 4th, 1981 production: the HGP-A well began operating, but damage to the plant forced a closure after a production period that had lasted a little over two months. Chemical analyses of the separated liquid and gas and some isotopic analyses of 34 S in the gas and in the sulphate are available (Thomas and others, 1984).

10 - December 11th to 22, 1981 flow test: chemical gas and separated liquid analyses are available (Baughman and others, 1985).

11 - In March, 1982, the power plant began operating. The nearly monthly chemical gas and separated liquid analyses are available only to the beginning of 1985 (Baughman and others, 1985). No other data referring to samples taken afterwards were collected.

__Synthesis of the hypotheses advanced in the literature__

It is extremely difficult to synthesize in a few lines all the material available in the literature about the HGP-A well. Many of the geochemical hypotheses elaborated during the initial stages of production (1976-1980) have been abandoned or changed as the well monitoring became more systematic (1981-1985). Furthermore, generally, the initial tests provide incomplete data about transient phenomena occurring each time the well was opened. Measurements taken after the power plant began operating reflect a practically stationary situation.

The most important aspects that have characterized the geochemical evolution of the HGP-A well and the hypotheses advanced by different authors to explain some of the variations observed, are summarized below.

As to the salinity of the separated liquid, in the initial phase of any well test it was always fairly less concentrated than at the end of the previous productions. At the end of each test salinity was always higher than in the previous one, ranging progressively between 3700 and 5800 mg/l in the 1976-1980 period (Kroopnick and others, 1978).

It also appears that for gases there was a significant per cent increase with respect to the total fluid occurred with each successive test. This increase, accompanied by an advancing higher concentration of H_2 and H_2S, was interpreted as a progressive approximation to the steady state composition of geothermal fluids (Thomas and Kroopnick, 1978).

Thanks to the greater regularity with which samplings were made once the HGP-A power plant began operating (end of 1982) it was possible to advance new hypotheses to justify the evolution of the gas chemistry occurred in the flow tests.

A front flash migration and the consequent heat mining inside the reservoir was hypothesized in order to explain both the radical change of the non-condensable gases concentration and the vapor fraction increase observed at the beginning of the tests and which were followed by their gradual stabilization (Thomas, 1982; Thomas and Sakay, 1983).

Until the end of 1989 HGP-A well remained constantly in production in stable conditions. On the basis of the samplings which were performed until the beginning of 1985, it was observed that the T.D.S. of the separated liquid had progressively increased to almost 17000 mg/l. The steam fraction, with a 12 bar a.b.s. pressure at the separator remained constant at between 43 and 44% (with a fluid enthalpy calculated at approximately 1650 KJ/kg).

The gas percentage in the steam (0.25% in weight), as well as the absolute concentrations and the relative proportions of the individual gases ($CO_2 \approx 53.7\%$, $H_2S \approx 40.2\%$, $N_2 \approx 5.6\%$, $H_2 \approx 0.5\%$ in average) (Baughman and others, 1985) also remained nearly stable.

The salinity increase found in the separated liquid was interpreted, at least in part, as a fluid withdrawal from a single phase-low salinity aquifer generating an expanding zone of boiling around the borehole (Thomas and Sakay, 1983). Thomas himself (1987), evidently taking into consideration the results of the KS-1, KS-1A and KS-2 wells, cased at a minimum depth of 2000 m and producing dry steam or excess enthalpy two-phases, reviewed some of the hypotheses stated above and advanced the following conclusions.

The progressive increase in the concentration of HGP-A's separated liquid (five times higher in 1985 than in the initial phase) could not depend on the migration of the production front because a corresponding proportional decrease in the separated liquid/vapor ratio had not been registered. Hence the single production level hypothesis was abandoned in favor of a model featuring two productive zones. The deeper of these zones would be entirely constituted by steam generated by total vaporization of an initially liquid phase; the second zone would be constituted by a two-phase. This theory would be consistent with the near stability in time of the content of the non-condensable gases (which decreased only by 10% keeping, however, the internal ratios between its various components). Production from a single level in which phenomena of vaporization and withdrawal affecting the front of the flash, in fact, would imply a fast decrease in the gas content in the vapor phase. This has not been observed.

The increase in the T.D.S. during the HGP-A production would hence be entirely due to the inflow of partially re-equilibrated seawater into the 1320 m production layer. The temperature of the fluid calculated with the Na-K-Ca geothermometer fell from 300°C in 1977 to 250°C in 1985. The value computed through the SiO_2 concentration, essentially unchanged, was stable at about 305°C.

Depending on whether it was applied to the total gas concentration present in the vapor or to the total discharge, the geothermometry of the gases showed, respectively, values of 380°C and 350°C.

The difference of the temperature values obtained from the geothermometers applied to the two different phases would confirm the presence of two productive levels. The deeper would be fed only by steam while the shallower would be fed by a mixed two-phase at a lower temperature.

5.4.5. TENTATIVE EVALUATION OF DEEP DYNAMIC PROCESSES

On the basis of the samples classification, none of the surface waterpoints, except for GTW-3, can be related to a deep geothermal component; it is therefore impossible to realize any deep circulation pattern. The GTW-3, however, could be contaminated by seawater and the possible geothermometric evaluation should be considered tentative and unreliable. The only available deep fluids come from deep wells and their reservoir temperatures have been directly measured. The combined use of different geothermometers still provides useful information.

The graphical method based on the triangular plot of Na/K and K/Mg ratios proposed by Giggenbach (Giggenbach, 1986) may be applied in order to understand the dynamic processes occurred in the deep horizons. In the triangular diagram thus plotted (Fig.9/An.A), most of the representative points of HGP-A, KS-1A and LANIPUNA 6 wells are near to the "full equilibrium" marked curve. Their Na, K and Mg analytical concentrations are, in fact, very close to the theoretical coexisting values for solutions that reach chemical equilibrium interacting with a mineral assemblage of volcanic origin at the temperatures measured in the reservoirs.

On the left side of this diagram, in terms of decreasing Na%, the Na/K equilibrium ratios for the stated assemblage at increasing temperatures are shown. Some of these theoretical values are represented as a set of isotherms originating from the right corner of the figure. In the lower side of the diagram the K/Mg equilibrium ratios are expressed in terms of Mg%, the corresponding theoretical values at different temperatures pick up a second set of isotherms originating from the top of the figure. The "full equilibrium line" is defined by the crossing of the same values isotherms.

To understand how the method works, it is necessary to explain briefly the most important characteristics of the two thermometric functions which, combined, form the diagram. The Na and K re-equilibration speed from the moment in which the fluids move away from the reservoir is decidedly lower than the Mg one.

So while the Na/K ratio retains the memory of the initial thermal situation for a long time, the K/Mg ratio not only changes rapidly as the thermodynamic conditions vary, but it is also highly sensitive to mixing phenomena with surface waters that contain Mg in far greater quantities than the geothermal fluids. As a consequence of all this, original deep waters (such as weirbox or separated liquid samples of producing geothermal wells) should normally lie near the "full" equilibrium line at temperatures very close to the measured ones in the reservoir. On the other hand, shallow fresh waters will be located in the right corner near to low K/Mg isotherms and far away from the full equilibrium line. Mixed fluids will be found in the central area of the diagram and the original reservoir temperatures should be inferred from the closest Na/K isotherms. The related K/Mg isotherms should point out the water-rock equilibration temperature at shallow levels after mixing.

After having pointed out the general characteristics of the diagram, the conclusions that can be gathered from the samples representative points will be illustrated. For a better comparative analysis, the data of the HGP-A well will be divided into two sets; the first includes downhole and separated water samples until year 1980, the second concerns the 1981-1985 period.

Analysis of the 1976-1981 HGP-A data

The downhole samples are shown in the diagram as black empty circles.

The F4 sample, collected at 1770 m immediately before December 1976, is the only anomalous sample being located in the lower left because of an unusual concentration of potassium. All the others are along line A. The correlation improves further if samples F39 and F40 of February 14th, 1977 are not considered. These samples, collected in the well at a level of 690 m, have a T.D.S. of about 8000 mg/l, that is approximately four times the value of the separated liquid in the corresponding production period. Besides the calcium and magnesium content is also at least ten times greater than in the separated liquid during the production. Therefore, in the productive aquifer at 690 m dynamic phenomena of inflow of more saline fluids may occur. Given the lack of data, it is not possible to advance definitely a hypothesis on the nature of these fluids but the concentration of the major elements suggest a contribution of partially evolved seawater.

The green points identify the separated liquids collected in 1976, 1977 and in 1978 (only one).

The purple points refer to separated liquid samples collected during the only flow test made in 1980 slightly after the recasing operation that excluded the 690 m productive layer. Their anomalous position could be attributed to the interaction between the deep fluid and the water used to kill the well. This action must have certainly caused dilution, local variation of the chemistry and, probably, thermal degradation.

During the 1981 test, which lasted for over two months, the fluid came from the two deepest productive levels with a composition slightly different from the previous one. The relative samples represented by blue points, are initially quite close to the full equilibrium curve for a temperature above 300°C and tend to move progressively upwards.

These latter points can be considered the most representative original composition of the HGP-A reservoir. It is noteworthy that they are very close to the KS-1A sample which is located on the full equilibrium curve at 320°C. It is reasonable that the same reservoir fluid feeds both the wells and that the different T.D.S. of the separated liquid depend on the different production mechanism (boiling in the reservoir).

Line A links the green triangles (sub-group A water points) at the right margin of the diagram (fresh water area) with black, green and blue samples. It can be concluded that until the recasing, the HGP-A well produced a mixed fluid only partially equilibrated. Two components can be singled out: deep reservoirs fluids (1960 m and 1370 m horizons) and a mixing of salty and meteoric water (690 m layer). The proportion among these components could clearly change depending on the conditions of production tests.

Analysis of the 1982-1989 HGP-A data

Of the analysis of the HGP-A data the samplings made in 1982 (once the power plant started to operate) are represented in yellow and in red for the period 1983-1985.

All these points line up unequivocally with the 1981 and KS-1A samples along line D.

From a chemical point of view, this trend can be interpreted as follows: in the reservoir that feeds HGP-A (without making any distinctions as to productive levels) there occurred a progressive and continuous inflow of water with a higher content of chlorides, calcium and magnesium than the fluid initially produced. The development of the process over time can be clearly seen in Fig. 10/An.A (diagram Cl/Mg vs. Cl). Starting in 1981 this ratio decreased tenfold while the chlorides increased five times. Very probably the chlorides present in greater and greater quantities, derive from partially evolved seawater. From an isotopic point of view, this hypothesis could be proved by simple stable isotope determinations. Its chemistry after standing in formation at high temperatures should be certainly different from fresh seawaters.

Qualitatively, the magnesium, calcium and sulphate content must have decreased considerably in comparison to the original composition, while potassium and silica, on the other hand, must have become more concentrated. From a quantitative point of view, the phenomenon depends on the temperature and on the progress of the reactions (whether or not the full equilibrium condition of the fluid has been reached). In any case, even if it is not possible to know the starting conditions at the time in which the mixing began, it is possible to make some considerations about the dynamics of the process by taking into account the different speeds at which some of the elements re-equilibrate. In fact, the analysis of the changes over time of the ratios among Na, K and Mg shows:

- (A) in the mixing area the solution gradually moves away from the initial thermal situation;
- (B) as far as the fastest modifying elements are concerned, the resulting fluid arrives in the feeding area of the well still in equilibrium with the reservoir conditions.

Assumption (A) is confirmed by the position of the points representing the fluid in the Giggenbach diagram. In relation to the Na/K geothermometric function, they move away progressively from the full equilibrium curve. The temperature, which was initially at 300°C, decreases constantly and reaches, for the samples collected in 1984-1985, a value of 250°C. Taking into consideration the properties of the function, it provides us information on the evolution of the fluid in the most peripheral areas of the reservoir, which are certainly affected by a progressive drainage of colder waters.

Assumption (B) is confirmed - at least apparently - by the fact that no thermal degradation of the produced fluid has been observed. Moreover, as to the K/Mg geothermometric function, the points representing the samples, well identified by line D, remained above the 300°C isotherm.

The analysis of a set of data, not yet published and including the subsequent period of the HGP-A production, shows that to the middle of 1986 the points representing the samples have continued to move upwards along line D of Fig.9/An.A.

The maximum value of the Na/K ratio around 16 (corresponding to a value of 48% of Na in the triangular diagram) may be interpreted as the period in which the inflow of sea-origin water in the reservoir has reached the highest point. Subsequently, the ratio Na/K has undergone a gradual decrease up to a value of around 11 in the last 1989 samples.

The ratio K/Mg has remained substantially the same, as was to be expected, on the basis of the considerations above.

This decrease of the ratio Na/K may be interpreted as the result of a gradual "sealing" of the fractures which interconnect the HGP-A reservoir with seawater. The circulation of the latter in a high temperature zone brings about the precipitation of an abundant quantity of minerals, mainly anhydrite.

Analysis of SiO_2 data

The re-equilibrium speed of SiO_2 at temperatures above 250°C, even though lower than that of Mg, is nevertheless very high. However, in the previous paragraph the use of function K/Mg has been preferred for the thermal evaluation of the fluid near the productive fractures.

There are two reasons which have mainly determined this choice:

- the analysis of K and Mg are far more numerous than those of SiO_2;
- the use of thermometric functions based on a ratio between two components is to be preferred in this case to one which refers to the absolute concentration of a single component.

When boiling occurs in a reservoir the concentrations of the deep liquid can differ from those computed by dividing the analytic concentrations of the separated liquid by the vapor fraction produced. An immediate use of these values could be misleading. Even in the case of HGP-A it cannot be excluded that the same phenomena which lie at the origin of excess enthalpy present in the KS-1A and KS-2 wells have occurred in reservoirs, although to a lesser degree.

Relations between GTW-3 and deep fluids data

The water points of sub-group (B) and of GTW-3 in particular, will be discussed. They are represented in the diagram (Fig.9/An.A) by red triangles. Except for KS-1A and LANIPUNA 6, which are very close to the full equilibrium curve, all of them are located in the shallow water area, slightly shifted from the sub-group (A) points. Excluding the GTW-3 points, all the remaining samples lie along line (C) which is the mixing line between fresh waters and seawater (SW).

The GTW-3 samples, which is the only water point not related to deep wells showing a deep component, lie along line (B). This is very close to the 240°C Na/K isotherm and links also the HGP-A downhole samples F39 and F40 (both sampled at 690 m). Three possible tentative hypotheses can be put forward, but none can be proven, given the lack of data:

- the GTW-3 deep component is the same fluid that fed the first productive layer of the HGP-A and whose estimated temperature should be 240°C;
- the GTW-3 is linked to deep fluids at an initial temperature of 250°C. During their circulation, these fluids disequilibrated and thereafter mixed with surface waters;
- the GTW-3 deep initial component might be ideally related to fluids with the same characteristics as those produced from deeper reservoirs of the HGP-A and KS-1A wells. The present composition should be linked to a contribution of seawater occurring in the last phase of circulation. This process can be graphically represented in the diagram with an increase in the slope of line A up to line B (rotation towards the SW point that represents seawater).

5.5. WELL DATA

In the KERZ area the drilling of deep exploratory wells started in 1975. The first well, called HGP-A, was commissioned by the University of Hawaii in order to verify the presence of a geothermal resource industrially exploitable. The well was productive and confirmed the presence of a high enthalpy geothermal reservoir.

Seven other wells (including one side-track well) were drilled in the 1980 - 1985 period afterwards. Three of them (KS-1, KS-1A, KS-2) were drilled by Puna Geothermal Venture (PGV) near HGP-A and were also found to be productive. The other four (Lanipuna 1, Lanipuna 1 S.T., Lanipuna 6 and Ashida 1) were drilled by BARNWELL Industries Inc. and were found to be not productive.

Some data relative to these wells and their location are shown respectively in Tab. 1/P2 and Plate 2. These wells are located in a rather small area; therefore the data collected from them cannot be significant for the entire area selected for the project.

The documentation examined is frequently incomplete and dishomogeneous: not all the well and production test reports are available.

Furthermore, the wells were drilled by different companies and not in the frame of a common deep exploration project for the area.

A synthesis of the main well data is given in the following paragraphs. The more reliable technical characteristics concerning stratigraphy, technical profile, temperature, pressure and geophysical logs for each well were selected and are shown in the figures collected in Annex B.

5.5.1. STRATIGRAPHY AND MINERALIZATION

The deep exploratory wells were drilled in a rather small area of the Lower East Rift Zone. Wells HGP-A, KS1, KS1-A, and KS2 are located in the middle of the rift, while wells Ashida 1, Lanipuna 1 and Lanipuna 6 are closer to its southern border.

The Lower East Rift Zone has witnessed volcanic activity in the course of the centuries. In recent times (1790, 1840, 1955 and 1960) there have been historical eruptions with voluminous emissions of tholeiitic basalt flows. The presence of many cinder cones, some of which are of hydromagmatic origin (Kapoho Cone), is evidence of a water-magma in-depth interaction.

The wells crossed a succession of tholeiitic subaerial basalts, of both "aa" and "Pahohoe" types, and shallow and deep submarine basalts. Locally, levels of tephra and intrusive bodies were also found, the former near the surface and the latter at greater depth.

The degree of vesiculation decreases with depth going on from subaerial to deep submarine basalts. In any case the vesicles do not have an important role in the permeability of the rocks because it is related principally to tensional structures, to the cooling contraction joint of the pillow lava and to discontinuities among lava flow units (Iovenitti and D'Olier, 1985; Macdonald, 1976). The zonation of hydrothermal and alteration minerals in all these wells is similar to those found in other geothermal fields in the world (Aumento and others, 1982; Aumento, 1985; Betancourt and Dominco, 1982; Bruni and others, 1983; Cavaretta and others, 1980; Elders and others, 1978; Fouillac and others, 1986; Goutierrez and Aumento, 1982; Kristmannsdottir, 1978 and 1982; Liguori and others, 1982).

The volcanic sequence crossed by the wells presents several layers with different alteration degree. Three specific types of alteration have been identified: deuteric, contact metamorphic and hydrothermal. The last two processes, which overprint the deuteric alteration cause the same alteration facies and hence are difficult to distinguish (Iovenitti and D'Olier, 1985).

In the HGP-A well three alteration zones have been identified under unaltered lavas; each is characterized by the predominance of certain minerals. From 675-1350 m, montmorillonite zone; from 1350-1894, chlorite zone; and from 1894 - b.h. actinolite zone. In the latter area epidote associated to quartz, chlorite, calcite, pyrite, anhydrite has also been identified, but it is not possible to determine whether it derives from contact metamorphism or from hydrothermal deposition even if the chemistry of the rocks and the temperatures are compatible with the stable epidote hydrothermal phase (Waibel, 1983). The increase in hydrothermal alteration and deposition starting at about 1350 m depth, in the chlorite zone might indicate the top part of a present-day or past convective system (Palmiter, 1976; Kitamura, 1980). The zone between approximately 1350-1959 m is slightly permeable and therefore could be a self sealing cover rock of an underlying reservoir (Kitamura, 1980).

The fluids produced have temperature ranges compatible with the distribution of the present stable hydrothermal minerals Waibel (1983), a phenomenon that has been encountered in other geothermal fields (e.g. Cerro Prieto, Iceland).

In the KS-1 and KS-2 wells the texture of the rocks beneath approximately 1350 m has been largely obliterated by alteration. Three zones have been identified in these wells: the first, unaltered, to 900 m (300 m deeper than in HGP-A), the second, a localized moderate alteration zone up to about 1200 m and lastly a zone with common moderate alteration with some layers of highly altered rock interspersed with occasional zones of fresh, unaltered rock. These fresh rocks might represent, at least in part, recent intrusions linked to the 1955 fissure vents and might constitute impermeable layers (Iovenitti and D'Olier, 1985).

Similar processes of hydrothermal alteration and deposition were found in the other wells. The presence of actinolite was not pointed out by all the drillings, but given the dishomogeneity of the available data it is not possible to determine whether this absence is due to an actual non-presence or to a failure to point it out. In the Ashida 1, Lanipuna 1 and Lanipuna 1/ST wells, the alteration is not continuous. While temperature is the main control factor of metamorphism, the intermittent alteration suggests the presence of fluid circulation. Anhydrite is greatly diffused in some wells (HGP-A) and apparently confined to single layers in others (Lanipuna 1, Ashida 1). It should deposit as a consequence of seawater circulation inside the fracture systems of the rift (Thomas, 1987).

It is difficult to establish a direct correlation between alteration-hydrothermal mineral suite and the present day permeability of the geothermal "reservoir", although some models of mineralogical association can constitute a good approach to the problem. The present-day permeability seems to be due almost exclusively to the presence of fractures which, though mineralized, can be continuously renewed both by magmatic intrusions and by tectonic stress releases.

For example, the highly fractured character of most of the cores of HGP-A is noteworthy. The renewal of the fractures can occur through the reactivation of the preceding fractures, as shown by the core of HGP-A, which present mineralized fractures, at times with evidence of successive tectonic movements (Macdonald, 1976).

5.5.2. GEOPHYSICAL WELL LOGGING

Geophysical well logs were performed in Ashida-1, KS-2, KS-1, and HGP-A, but data are available only for the last two.

The original logs of these wells are shown in Fig. 2/An.B and Fig. 4/An.B respectively and have been slightly filtered in order to make the representation less confusing.

The logs, which for both wells investigated only one part of the hole, were: Gamma Ray (GR), Self Potential (SP), Neutron (SNP) as well as some resistivity logs, both conventional Short and Long Normal (SN and LN) and Induction (ILD) logs.

Considering the kind of logs carried out, their aim was basically to find porosity and permeability variations in reference to the texture variations of the rock.

In HGP-A, which has its bottom hole at 1968 m, only GR and SNP were carried out from the well head at a depth of approximately 1050 m, while SP and resistivity logs were performed only between 680 and 1050 m.

From the SNP it is possible to observe that the values of the Api units increase progressively with depth. This indicates a consistent general diminution of porosity and permeability. In particular, the high porosity of the shallow basaltic formations to a depth of about 300 m is evident. A further increase in the SNP values and a slight increase in the values of the GR curves is observed at about 680 m. The magnitude of this increase might be exaggerated passing from the casing section to the open hole, but the identification of denser formations, characterized by a dense alteration of more or less permeable levels, is nevertheless valid. The greatest GR values might indicate a concentration of radioactive minerals deposited along such fractures by geothermal fluids.

An integrated interpretation of the available logs was carried out for the interval between 680 and 1050 m, (Rudman, 1978). In particular, by interpreting the negative deflections of the SP curve as indicating permeable horizons and the higher resistivity values as indicating dense and not very porous formations, the 945-975 m and the 1010-1035 m intervals stand out for their higher permeability.

Information about the deeper part of the well, which is, from a geothermal point of view, more important, is unfortunately lacking.

For a calibration of the surface geophysical prospecting the data are also insufficient. In fact, the only element that can be identified, is the passage, at about 940 m, from a layer with a 10 ohm*m resistivity to a lower layer with a 2-5 ohm*m resistivity.

In the KS-1 well the only logs performed along a significant interval (275-1225 m) are SP and resistivity. GR and neutron logs were carried out only for the first 250 m.

The most significant characteristic is the high permeability marked by an abrupt negative deflection of the SP curve and by a net decrease in resistivity at the depth of about 450 m.

Also in this case there are no data on the deepest and most important sector of the well, while the only calibration component for the surface geoelectrical data is furnished by the induction log. In particular, the first resistive stratum reaches a depth of about 450 m; it is characterized by resistivity values ranging from 100-300 ohm. m. The resistivity of a second conductive stratum seems to decrease with depth (from about 50 to 2-5 ohm*m). This can probably be associated both to the presence of argillification and to an increase of permeability with geothermal fluid circulation. This might occur to a depth of 750-800 m.

5.5.3. TECHNICAL CHARACTERISTICS OF THE WELLS

Several temperature surveys have been carried out inside the wells in a static condition and with different stand-by times. Unfortunately, the well bottom temperature recovery survays performed soon after the stopping of the drilling, have not been properly carried out. The pressure profiles in static conditions are available only for HGP-A, KS-1 and KS-2. Injection and/or production tests were made only for some wells.

<u>HGP-A Well</u>

The HGP-A well was drilled between December 10, 1975 and June 1976, and reached a final depth of 1968 m.

During drilling, absorptions and total loss circulation (T.L.C.) occurred in the shallower layer, to a depth of about 100 m.

Afterwards, and particularly after the 9" 5/8 casing setting at a depth of 675 m, the drilling continued using mud as drilling fluid and very limited absorptions were detected.

At the end of the drilling and after the setting of the 7" slotted liner, the mud was replaced with water. The injection tests carried out on June 6 and 7 highlighted the very low injectivity of the well as can be seen from the following data:

Injection flow rate	Well-head pressure
77 m^3/h	41 bar
24.5 m^3/h	27.5 bar

After these tests the well was kept in stand-by for the warming-up. Flow tests with air lift were carried out between June 22-24 and on July 19 and 29, 1976.

Several T and P surveys were made during the warming-up phase and after the flow tests. The results of these surveys are shown, along with other characteristic parameters of the well, in Fig. 1/An.B.

The temperature profile shows that under 1065 m temperatures range between 300° and 350°C, and that the maximum temperature seems to reach about 358°C at 1968 m.

The surveys carried out confirm the presence of a liquid dominated reservoir in which the temperature distribution with depth is close to the boiling point.

A silencer/separator unit was then installed on the well and numerous flow tests were performed in order to characterize, from the chemical-physical point of view, the fluids produced and to determine the hydraulic characteristics of the formation.

In the course of the various production tests performed in the period from November 1976 to March 1977, an increase in the flow of the well was observed. This phenomenon is to be attributed to the progressive cleansing of the fractured productive layers by the drilling mud which was solidified by the high temperature. According to these tests the possible electrical output was estimated to be between 3.1 and 3.5 MWe.

The analysis of the pressure drawdown and build-up tests confirms the presence of a low formation permeability and of damage to the well caused by the drilling mud. The various tests agree on the permeability-thickness values of $0.3 \div 0.4 \cdot 10^{-12}$ m^3 and skin values of $4 \div 6$.

The temperature and pressure profiles carried out during the flow tests show the presence of a two-phase flow inside the well. Due to the low permeability and high temperature the flash occurs inside the formation.

The most productive zones are located at around 1310 and 1830 m; in addition, a productive zone is located immediately under the shoe of the 9" 5/8 casing (675 m).

In September 1979, after the well was damaged during the flow tests, a 7" casing was set with the shoe at 915m. The new technical profile is shown in Fig. 1/An.B.

In June 1981 the 3 MW experimental unit construction was completed and from that time on the well has remained in operation for the production of electric energy with only brief shut-in periods.

In December 1989, once the experimentation period was over, the well was closed and the power station shut down.

During the operation of the power plant, the production characteristics remained practically constant with the following mean values:

- Total Flow Rate : 50 t/h
- Flowing well-head pressure : 12 bar
- Enthalpy : 1650 KJ/Kg

The well has confirmed the presence of an industrially exploitable geothermal system but unfortunately, despite the long production period and the presence of other wells drilled in the same area, interference tests have not been carried out and there is no information about the trend of the reservoir pressure during the fluid production.

KS-1 Well

The well was completed on November 12, 1981 at a total depth of 2222 m.

The technical, stratigraphic and temperature profiles are shown in Fig. 3/An.B.

In the shallower part of the well high absorption and diffuse T.L.C. to a depth of 280 m have been observed.

After the setting of the 13 3/8" casing, limited absorptions were observed starting from a depth of 335 m; these progressively increased from a depth of 2170 m, and reached the value of 38 m^3/h at well bottom.

A temperature survey performed after 20 days of warming-up shows a uniform temperature of about 320°C in the stretch underlying the 9" 5/8 casing. The uniformity of the temperature indicates the presence of an interzonal flow.

The well was found to be productive; the following production parameters were determined in the course of a series of tests performed between August 11 and 28, 1982:

- Flow rate: 32.5 t/h of dry steam
- Flowing well head pressure: 8.2 bar
- Enthalpy: 2756 kJ/kg

Pressure surveys in static conditions are not available. On the basis of the information collected from other wells drilled in the same area, the presence of a liquid dominated system with a temperature distribution close to boiling values can be assumed. The dry steam production would derive from flashing inside the formation.

After the production tests the casing revealed damages at several points. The well was therefore cemented in and considered unexploitable.

KS-1A Well

The well was completed on September 3, 1985. The technical and stratigraphic profiles are shown in Fig. 5/An.B along with some temperature and pressure profiles carried out in static conditions after the production tests.

The well, which was drilled by the Diamond Shamrock's Thermal Power Co., reaches a maximum depth of 1983 m. In the literature collected and analyzed there are no data concerning the stratigraphy, the absorption and/or circulation loss zones or injection tests that might permit the identification of fractured zones.

However, the well was found to be productive. The production data, obtained during a test on October 31, 1985 are as follows:

- Total flow rate: 36 t/h
- Flowing well-head pressure: 11.5 bar
- Enthalpy: 2417 KJ/Kg

The well produces a two-phase mixture with high enthalpy, which at a separator pressure of 10.3 bar g, has a steam quality of 82%.

The temperature surveys carried out in static conditions after the production test show nearly constant temperature ranging from 330°C to 340°C in the stretch with slotted liner.

The pressure surveys carried out in static conditions confirm the presence of a water dominated system.

The uniform temperature in the stretch with slotted liner should be attributed to convective circulations.

KS-2 Well

The well was completed on April 2, 1982 and reached a maximum depth of 2440 m. The technical and stratigraphical profiles and the P and T surveys carried out in static conditions are shown in Fig. 6/An.B.

During drilling highly fractured zones with T.L.C. between 98 and 402 m were identified in the first stretch. Afterwards drilling was continued with absorptions varying between 1.5 and 13 m^3/h, which at about 2220 m increased to approximately 45 m^3/h.

The well was found to be productive and during the flow tests performed between July 28 and August 2, 1982 the following production characteristics were determined:

- Flow rate: 15 t/h of dry steam
- Flowing Well Head Pressure: 13 bar
- Enthalpy: 2780 kJ/kg about

When the flowing well head pressure decreases to values below 10 bar the well produces a two-phase mixture.

The static well temperature surveys reveal a uniform distribution in the stretch with slotted liner, with values between 330-340°C, as in the wells KS-1 and KS-1A.

The static well pressure surveys confirm the presence of a liquid-dominated reservoir. The reservoir pressure measured at the same absolute depth is very close to the values measured inside the well HGP-A. The well had casing damage and was cemented in. The well could be reworked.

__Lanipuna 1 well__

It was drilled between February 9 and May 26, 1981 by Barnwell Industries Inc.

The well highlighted the presence of fractured zones in the first stretch. No circulation losses or absorptions were found to the total depth of the well (2257 m) after the 13" 3/8 casing with shoe at 311 m was placed.

During the drilling, several temperature surveys were made. The most significant are shown in Fig. 7/An.B along with the technical and stratigraphic profiles.

A conductive thermal trend, under the 13" 3/8 casing was identified by the surveys, except for the stretch between 1700 and 1920 m in which the temperature is practically constant. This uniform temperature might indicate the presence of a permeable zone near the well, with temperatures probably around 250°C.

Below 1920 m, the very high gradients (about 4°C/10 m) make it possible to extrapolate a temperature value of about 400°C at well bottom.

On April 20 and 21, with the well bottom at 2133 m, air-lift flow tests were carried out. The flow was not substained without the aid of compressed air and the tests revealed an increase in salinity of the fluid produced, from 10 to more than 10.000 ppm, indicating the entry of some formation fluids.

The temperature distribution of the well in static conditions (Fig. 7/An.B) suggests that the entry of formation fluids occurs in the stretch between 1700-1920 m.

The injection tests carried out after the production tests, revealed the low injectivity of the formation:

Injection flow rate $[m^3/h]$	Well-head pressure [bar]	Injection time [hour]
38	42	2
38.6	42	1
23.8	31.5	5

Unfortunately no temperature logs, which would have made it possible to identify the absorption zones, were performed during the injection tests.

Temperature surveys carried out after the injection tests revealed temperature inversions indicating a smaller entry of water just below the shoe of the 9" 5/8 casing (1067 m), and prevalently at a depth of about 1200 m.

Considering the lack of permeability found at these depths during the drilling, the absorptions in the course of injection tests must have occurred because of fractures artificially generated by the high injection pressure.

After these flow and injection tests, the well was deepened to 2557 m. No absorption zones were found and no other tests were carried out.

Lanipuna 1 S.T. Well

The Lanipuna 1 S.T. is a directionally drilled hole which departs from the Lanipuna 1 geothermal exploration well at a depth of 1088 m with a N 20° E direction. The drilling operations were performed between May 17 and June 19, 1983. The total depth of the directionally drilled hole is 1970 m (vertical depth, 1911 m); the bottom hole is located at a distance of 278 m in a 22° N direction from the drill site.

No absorptions were found during the drilling except for a very small absorption of only 3000 l registered at a depth of 1295 m.

This sidetrack well did not prove, therefore, the existence of the permeable layers at depths between 1700-

1920 m, hypothesized on the basis of the Lanipuna 1 temperature profiles.

The temperature measurements in static condition taken at the end of the drilling (Fig. 8/An.B) show a thermal profile that is considerably different from that of the Lanipuna 1 well. The measurements revealed a strong temperature inversion which cannot be correlated to induced cooling phenomena since no evident absorption occurred during the drilling phase.

The maximum temperature measured was of about 220°C at a depth of 1650 m.

It should be noted that the thermal gradient flatting of the Lanipuna 1 and the thermal inversion of Lanipuna 1 S.T. are both found approximately in the same stretch (1700-1900 m).

At the end of the drilling the mud was displaced from the hole with water and the well was kept in stand-by for a month. During this month static well temperature surveys, the most significant of which are shown in Fig.8/An.B, were carried out.

In the period between July 27 and July 29 the well was unloaded with compressed air, first to a depth of 610 m and later to a depth of 915 m.

During the unloading tests a maximum of 5-6 m^3/h of fluid entry was observed.

The well was, therefore, considered unproductive and completed with a cement plug.

Lanipuna 6 Well

The well was drilled for exploration purposes by Barnwell Industries Inc. in the period between February 22 and June 1, 1984.

The technical profile of the well, in fact, is more typical of an exploration well than of a production well.

The drilling highlighted the presence of a practically impermeable formation up to a depth of 1309 m, where a T.L.C. was found. The drilling continued to a depth of 1341 m without return of circulation fluids.

Several attempts were then made to plug the fractured zone with cement in order to continue drilling. These attempts were unsuccessful; furthermore in the course of a cement plug redrilling, the well went in spontaneous deviation starting from a depth of 445 m.

The new stretch of well confirmed the presence of an impermeable formation up to 1309 m where T.L.C. was found again. Once again, unsuccessful attempts were made to plug the fractured layers.

The drilling continued without circulation to a depth of 1510 m.

During the drilling many temperature surveys with different stand-by times were made.

The most significant of these surveys are shown in Fig.9/An.B. The temperature measurements highlight a conductive thermal regimen from about 600 m to a depth of about 1300 m, where a maximum temperature of 138°C was measured after 34 days of stand-by.

Beyond this depth the temperature drops abruptly. Considering the drilling history, this phenomenon could be due to the cooling induced by the absorption of the drilling fluid in a fractured zone.

Ashida 1 Well

In 1980 Barnwell Industries Inc. drilled the Ashida 1 well. The maximum depth of this well is 2530 m; its technical and stratigraphic profiles are shown in Fig. 10/An.B.

In the shallower stretch numerous absorptions and lost circulations were identified.

After the setting 13" 3/8 casing with shoe at 373 m, drilling continued without further absorptions. The well, therefore, was found to be unproductive.

Several temperature measurements were made in the course of the drilling and afterwards, the most significant of which have been selected to identify the temperature formation distribution.

Beyond the first and highly permeable stretch, the temperatures of which are $\leq$ 50°C up to about 1000 m, these measurements revealed a decidedly conductive thermal regimen with an average gradient of about 4.5°C/10 m. The maximum temperature measured at well bottom, after 25 hours of stand-by, was 287°C.

However, on the basis of the different thermal surveys made with longer stand-by times, the stabilized well bottom temperature can be assumed to be about 300°C.

6. INTEGRATED ANALYSIS OF THE DATA

The integrated analysis of the data available permits a reconstruction of both the regional model of the KERZ and of the geothermal model of the Lower East Rift Zone.

The former has been amply defined by various authors and undoubtedly represents the most advanced and scientifically valid interpretation.

The geothermal model of the LERZ needs a critical evaluation because of the lack of data available, from both a quality and quantity point of view.

6.1. REGIONAL FEATURES OF THE KERZ

The area of the project is situated entirely in the East Rift Zone of the Kilauea Volcano.

The rift is about 45 km long structure from the summit of the volcano to Cape Kumukahi and then continues beneath the sea.

The northern part of the rift is generally inactive while the southern part is characterized by continuous phenomena of rifting.

The outcropping part of the volcano is formed for the most part of vesicular flows of tholeiitic, olivin tholeiitic and picritic basalts, to which discontinuous pyroclastic deposits, often palagonitized, are associated. The deep wells have crossed these flows for hundreds of meters lying on other ones deposited in shallow sea. These flows are vesicular too, often palagonitized and widely altered. They lie on top of other deeper submarine flows which are completely void of vesiculation.

The KERZ is characterized on the surface by a series of fractures, faults, feeding fissures, scoria cones, pit craters, and lava shields. Many of these volcanic features were active during the last decades and centuries.

The Upper ERZ, in particular, is characterized by en echelon fissure vents, while the Middle ERZ presents swarms of numerous fractures and faults in the northern part predominantly. The Lower ERZ has fewer faults and fractures and considerably longer fissure vents.

The northern part of the Middle ERZ, where there is a higher number of fractures, presents flows of more than 250 to 350 years old compared to other zones of the rift which are characterized by more recent volcanic activity. Therefore, the superficial fractures in these zones could be covered by subsequent volcanic effusions.

A series of dikes which stretch across the rift lengthwise form the **Dike Complex** and have permitted the lateral migration of the magma from the main summit chamber (Walker, 1988).

The correlation between hypocentral distribution of earthquakes and magma activity has allowed to delineate the preferential paths of the magma ascent and intrusion. It is assumed that the magma chamber of the Kilauea is located at a depth of neutral buoyancy (Ryan, 1988) between the less dense shallow flows and the older and deeper ones; the latter are denser as a result of magmatic injections (dikes or intrusions). This chamber, therefore, is gravitationally stable (Walker, 1988) and can be located within a depth of 6 km (Koyanagi and others, 1974; Klein and others, 1987).

The integrated interpretation of various geophysical data (Ryan, 1988) suggests the existence of a second level of neutral buoyancy at a depth of 6-12 km. This second level corresponds with a denser, ultrabasic, picritic magma. Intrusions stretch towards the KERZ from this deep magma chamber also and being denser, stabilize at the same depth of the reservoir constituting a **Deeper Rift System.**

The high pressures reached by this ultramafic magma force the masses of the South Flank and produce the great Hawaiian earthquakes. It is interesting to note that after the severe Kalapana earthquake in 1975 (M=7.2) the seismicity at intermediate depth in relation to the summit zone of the Kilauea decreased whereas there was a considerable increase in the South Flank and in the LERZ (Klein and others, 1987).

The structure of the KERZ is also shown both by magnetic and gravimetric surveys. In particular, deep rift systems contribute to the gravity deep anomalies (Woollard, 1951; Strange and others, 1965; Kinoshita and others, 1977) while the most significative magnetic features (Flanigan and others, 1986) are the strong linear magnetic gradients and the dipolar magnetic anomalies, which, with E-W trend, correlate with the alignment of important volcanic-tectonic elements. These anomalies are linked to intrusive basaltic bodies, by now cool, of the Dike Complex.

This complex has been outlined also by the gravimetric modelling (Broyles, 1977) carried out on the basis of a detailed gravimetric survey (Furumoto and others, 1976). The best fitting, in fact, is obtained with bodies situated at a depth of 2 to 6 km and with a density which corresponds with basaltic-olivinic type intrusive rocks.

Furthermore, seismic refraction surveys evidanced a fast body in connection with the Dike Complex at a depth of 5-6 km which, in accordance with gravimetric data, has a lateral extension of 12-19 km (Furumoto, 1978). The effective extension of the Dike Complex is, therefore, greater than the surface width of the rift which is of the order of 5 km.

The South Flank of the KERZ is currently active and characterized by continuous intrusions of dikes. The intrusions are the direct cause of the southward displacement of the South Flank of the KERZ which is not supported laterally. The North Flank of the KERZ, instead, is relatively stable because of the buttressing effect of the huge mass of Mauna Loa. Furthermore, the gravitational stress field imposed by the huge mass of the Mauna Loa, on which the Kilauea is built, determines the gravitational assessment of the free slope of the whole mass and favors its displacement southwards, externally from the center of growth of the volcanic sequence (Swanson and others, 1976).

The South Flank of the volcanic edifice is thus broken up into deep tectonic blocks which fracture internally when forced towards the sea. Also this phenomenon determines intense seismic activity in the South Flank of Kilauea. The faults of the Hilina System, interpreted as deep listric faults, define the limits of these tectonic blocks on the surface (Swanson and others, 1976). This fault system dissects all the southern part of the volcano; its surface effect decreases towards Kalapana with probable extensions into the sea. N-E of Kalapana the faults of the South Flank are no longer evident on the surface: they are either covered by recent extrusions, or they do not exist as is suggested by the distribution of the epicentres which shows a strong attenuation of seismic activity. The few epicentres in the Lower ERZ are distributed near the rift zone and are therefore more probably connected to volcanic activity.

The structure of the Dike Complex of the KERZ has been hypothesized by means of a comparative study carried out between the Dike Complex of the Koolau and Kilauea volcanos (Walker, 1986). The rift is injected by dikes which stretch along the sides or the preceding ones and have a density of 50-100 dikes for 100 m.

There are fewer injections in the rift along its flanks and its shallow part; in these parts, being younger, there is only evidence of more recent intrusions. The dikes, with an average width of 65 cm, are associated in clusters up to a width of 20 m lengthwise the rift; they also form two complementary sets with opposite slopes of 65°-85°.

The non-verticality of the dikes is to be attributed to the anisotropies caused by the shape of the volcanic edifice and by its tendency to spread laterally. The development and distribution of the dikes are a consequence of their distance from the central reservoir. They are, in particular, deeper and more numerous near the summit zone, whereas they become shallower and less numerous in the more distal ones (Walker, 1988). This would be confirmed by the lower seismicity which characterizes the LERZ.

The injections of dikes without eruptions are more frequent than fissure eruptions (Dzurisin and others, 1984). This is typical of a magma which, once it has reached its gravitational equilibrium between hosting rocks of the same density, tends mostly to rest rather than to extrude (Walker, 1988). This characteristic has allowed the settling of magmatic stocks in the rift which remain molten for such a long period to undergo differentiations and to feed successive eruptions.

The possible existence of magma chambers has been shown on the basis of petrographic characteristics of the material erupted. In the LERZ, magma chambers are associated with the vents of HEIHEIAHULU, KALIU and PUULENA-HONUAULA (Moore, 1983).

All these volcanic features, however, are located in correspondence of the shallow body causing the lengthened magnetic anomalies of the rift. The magma chambers, therefore, must be deeper than the magnetic bodies themselves which, according to the proposed models (Flanigan and others, 1986), should reach at least a depth of 2 to 3 km.

Among the various circular magnetic anomalies which characterize the rift, some present an apparent inverse polarization which is certainly caused by topographic effects. On the other hand, the anomaly near the Kapoho crater could be interpreted as a non-magnetic structure associated with a shallow magma chamber with a higher temperature, therefore, than the Curie point (Flanigan and Long, 1987).

Some of the above considerations lead to a differentiation between the Upper/Middle part and the Lower part of the KERZ.

In fact, the interrelated effect of buttress and of the gravitational stress field on the rift development decreases progressively towards the East. As a consequence, while the Upper and the Middle ERZ are active only on the southern flank, the LERZ can be active and free to move on both flanks (Rudman and Epp, 1983).

This assumption could be confirmed by the different distribution of the emission fissures which in the LERZ are active on both flanks.

Further east in the LERZ, some geophysical evidences have pointed out a NNW-SSE transverse offset in the Puu Honuaula vent area, where the HGP-A, Kapoho and Lanipuna 1 wells have been drilled. A detailed gravimetric survey (Furumoto and others, 1976) first showed the breakdown of the gravimetrically positive structure with displacement north of its easternmost part. This breakdown could be interpreted as the effect of the sharp decrease of the shallow dikes intensity (Furumoto, 1978). Subsequently, an aeromagnetometric survey at low level highlighted, in the same area, an interruption of the body which causes the magnetic anomalies stretching towards the rift. (Flanigan and others, 1986). An aero-electromagnetic survey (VLF) evidenced an electrical conductive surface anomaly in the same area (Flanigan and others, 1986).

This transverse offset on the surface seems to divide the LERZ in two zones with a different distribution of superficial fractures and of fissure vents. Despite the lack of evidence on the surface, the transverse offset has been drawn on the reconnaissance geologic map of the Kilauea Volcano (Holcomb, 1987) as a right strike-slip fault. A movement towards the left, however, would be more coherent with Rudman and Epp's findings (1983), at least in its northernmost part.

Therefore, the LERZ seems to be structurally differentiated from the MERZ and also shows transverse discontinuity clues. These could represent releases belts between zones with different freedom of movement.

No direct data exist which can clearly identify the extent of the thermal anomaly. The shallow and deep structural model of the KERZ has highlighted a wide intrusive and extrusive magmatic activity which periodically dissipates considerable heat. The Dike System is therefore surely responsible for a regional geothermal anomaly in the KERZ.

When present, the satellite magmatic chambers can increase the regional heat flow. More locally, geothermal anomalies could be further increased by hydrothermal systems which facilitate the heat transfer by convection.

A geothermal assessment has been carried out on the basis of these data and of the whole geological, geophysical and geochemical information; it has evidenced that there are high possibilities of finding high temperature geothermal resources in the rift (Thomas and others, 1979; Thomas, 1986).

As far as anomalies due to convective systems are concerned, at present nothing can be said about their distribution; but, it can be stated with some confidence that there exist geothermal anomalies due to both the Dike System and the magmatic chambers.

6.2. GEOTHERMAL MODEL OF THE LOWER EAST RIFT

The well data available, which refer to the alterations, mineralizations, distribution and extent of the T.L.C. zones, formation temperatures, and to the geophysical logs were not collected systematically nor with analogous methods.

The drillings were, indeed, carried out by different companies within restricted mining leases and without common project coordination.

For this reason, on the basis of stratigraphic, mineralogical, hydrologic and thermal information available, it is not possible to individualize reliable guide levels, nor to correlate the productive zones of each well.

Only the data which refer to alterations and mineralizations of the Lanipuna 1, the HGP-A, and the KS-1/KS-1A wells, although not homogeneous, could indicate a correlation between the mineralized zones.

Figure B correlates the above-named wells, along a South-North section, and shows the approximate attitude of the principal mineralized zones which, from top to bottom, may be divided into <u>Montmorillonite Zone</u>, <u>Chlorite Zone</u>, <u>Epidote and Actinolite Zone</u>. The few data available indicate that both the shallow <u>Slightly altered Zone</u> and the above mentioned mineralized zones do not present a sub-horizontal attitude.

Among the geophysical characteristics, the electrical resistivity parameter can better outline variations connected with altered zones and hydrothermal circulation systems.

Unfortunately, the data available, although not few for the LERZ, are derived from diverse studies and surveys with different depth of investigation. It is therefore very difficult to give a reliable characterization as regards resistivity and thickness of the layers. Furthermore, there are very few electrical logs performed in wells and only for brief intervals. However, it is possible to attempt an electrostratigraphic characterization of the zones with different alteration degree (Fig.B).

The slightly altered zone can be electrically divided into a highly resistant top layer and into an underlying complex of medium resistivity. The former corresponds to highly porous terrains on the surface extending in depth to the water level which corresponds more or less with the sea-level. Above sea-level, all the surface data (EM, DD, VES) show resistivity values ≥ 1000 ohm*m.

On the other hand, as far as the underlying layer is concerned the data do not agree. Both the log carried out in the KS-1 well and a number of electromagnetic surveys (Skokan, 1974; Kauahikaua and Mattice, 1981) indicate a resistivity values of between 50-200 ohm*m as far deep as 500-1000 m. This however seems to contradict other EM surveys (Kauahikaua and Klein, 1977; Kauahikaua, 1981) where a highly conductive stratum (2-4 ohm*m) with thicknesses between 500-1000 m, under the shallow resistivity layer, has been detected.

Higher conductive areas, within this slightly altered zone, are certainly possible in correspondence with mixing between deep warm waters and surface waters, but should be very much localized. Therefore, the aforenamed conductive structure (2-4 ohm*m) might derive from an improbable correlation between survey sites remarkably distant from one another.

On the whole, therefore, all the slightly altered zone could constitute an electrically resistant body.

On the other hand, a conductive complex can be certainly associated with the montmorillonite and chlorite zones. The geophysic logs confirm, in particular, a correspondence between a stratum with about 10 ohm*m resistivity and the montmorillonite zone.

The underlying chlorite zone seems to be even more conductive with resistivity values which do not exceed 3 ohm*m. This indication is, however, given only by the log carried out in the HGP-A well and at a depth range of only 200 m. However, the possibility of a higher conductive and deeper body is suggested even in EM studies (Kauahikaua, 1981).

Finally, the epidote and actinolite zone should certainly be associated with the deep resistive basement with resistivity values which, as indicated by the most recent EM studies (Kauahikaua and Mattice, 1981) should exceed by 100 ohm*m.

Field characterizations, on the basis of other geophysical parameters are not possible as a consequence of the regionality of the surveys carried out (gravity) or because of their exiguity (seismic profiles).

Figure B also shows the temperature distribution considered as the most indicative of the undisturbed formation values. The thermal log of 5/1/1976, in particular, carried out during drilling 3 days after the stopping of mud circulation in the HGP-A well, has been considered the most significant. In fact, the temperature measurements carried out subsequently are influenced by phenomena of interzonal flow between the deep productive level and the shallower permeable ones. The temperature values of 5/1/1976 have been shifted so that the maximum temperature at 1400 m coincides roughly with the temperature of the fluid produced subsequently (test of 9/26/1976).

The Lanipuna 1 well shows a conductive gradient regime which is more or less constant except for the 1700-1900 m interval, where the temperature remains constant (250°C). This well, however, has proven to be practically unproductive.

In the HGP-A and KS-1 wells, instead, underneath a conductive zone, the geothermal gradient profile takes the usual shape of a convective regime. In both the wells, this zone of thermal flatting highlights the presence of a geothermal reservoir and corresponds with the effective production levels.

The temperatures distribution presents numerous lateral variations. The wells drilled in the Kapoho area, in particular, have shown a hydrothermal system with temperatures between 330-350°C at a depth of 1500-2000 m. At these same depths, instead, the wells drilled on the southern flank of the rift turned out non-productive with temperatures between 200-250°C.

Even the geochemical data, although limited, confirm the presence of at least two systems with distinctly different thermal characteristics. In fact, in the area of productive wells the analyses of fluids show congruent Na/K ratios with temperatures at around 320°C, whereas in the southern marginal areas (Lanipuna wells) these ratios indicate temperatures ranging between 180-250°C.

Furthermore, whereas the Lanipuna 1 well has a prevalently conductive regime with a bottom hole temperature (2557 m) of presumably 400°C, a thermal inversion with temperatures between 220 and 160°C at 1700-1900 m respectively was registered in the Lanipuna 1 side track well.

Such dissimilar temperatures, in wells so near to each other, testify a discontinuous hydrothermal system which instead is certainly continuous in the restricted area of Kapoho.

The geothermal system, therefore, appears to be connected exclusively to a system of fractures lengthwise the rift, the distribution of which is discontinuous and not homogeneous.

This geothermal characterization, however, cannot be applied to the whole rift area. The wells, in fact, were not drilled like exploratory wells of the Geothermal/Cable Project and are not, therefore, distributed uniformly in the selected subzones.

Although the reconstructed distribution of hydrothermal minerals is approximate, it is possible, nevertheless, to note a correspondence between the Montmorillonite Zone and the intervals at a higher geothermal gradient, for the HGP-A and KS-1 wells in particular. In addition, taking into consideration other geothermal fields in analogous volcanic environments where the role of cover and reservoir is strictly linked to phenomena of hydrothermal and thermometamorphic alterations, a characterization of the LERZ geothermal system may be inferred.

Starting from the bottom, the Epidote and Actinolite Zone seems to correspond to a reservoir with secondary permeability connected to a fractures system lengthwise the rift.

The Chlorite Zone may be considered the upper part of a convective system of high temperature; this correlates with the intervals characterized by a flatting of the thermal gradient. Locally, however, intense phenomena of self-sealing can drastically reduce permeability.

The reservoir should be confined towards the top by a layer with a variable thickness. This layer corresponds with the widely altered and mineralized Montmorillonite Zone and can, as a whole, act as an efficient cover for the deep geothermal system. Its top can vary between a depth of 600 and 900 m below ground level.

Even this latter zone, however, in correspondence with intrusions by dikes or by the ascent and emission of magma, may be partially permeable. Locally fissures and fractures can interconnect the deep reservoir with the shallow hydrogeological system. This corresponds with the slightly altered zone, which is very porous and permeable.

Its high permeability derives both from primary characteristics (vesiculation, cinder, clinker, tube, etc.) and from fractures.

In this shallow hydrogeologic system the abundant rain waters permeate rapidly in depth. They are impounded between the dikes and are drained away mainly lengthwise the rift. Locally circulations transverse to the rift can, however, occur towards the basal waters of the South Flank Zone.

In the entire KERZ intense drainage and a considerable altitude difference between the topographic surface and the water table, which practically coincides with the sea level, hide any thermal manifestations (gas, fumarole, thermal spring).

The shallow hydrological system is characterized by an intense mixing with seawater and its sea level temperature, generally ranging between 30 and 55°C, is linked both to conductive and convective phenomena. The former is due to local heat sources (dikes), the latter to deep thermal fluids uprising along fractures. This is the case of the GTW-3 where the chemical determinations have identified inflows of deep fluids. A deep inflow in the other water points, being strongly contaminated by seawater, cannot be pointed out by chemistry. The availability of isotopic and trace component data could allow a better discrimination among meteoric, deep water and seawater mixings.

Furthermore, the few available measurements and their distribution do not allow to estimate an average temperature value for the shallow aquifer. As a consequence, it is impossible to state a tentative heat and mass balance in order to evaluate the real extent of the convective phenomenon.

The entire shallow and deep hydrothermal system is characterized by a substantial pressures equilibrium which follows a hydrostatic-type regime.

As far as the deepest thermal aquifer is concerned, the chemical composition of the liquid samples of the HGP-A, KS-1A and KS-2 wells suggests that the wells are fed by the same reservoir or at least by reservoirs with similar characteristics.

On the other hand, the thermodynamic characteristics of the fluids as a whole are very different. HGP-A produces a mixing of liquid and vapor with a slight excess of enthalpy compared with the liquid enthalpy at the temperature measured in the fracture. The other wells, instead, were characterized by a noticeable excess of enthalpy already in the initial tests, and subsequently the enthalpy increased up to the threshold of the dry vapor. These differences are certainly due to the different processes which regulate the boiling phenomena in the reservoir.

During its production the HGP-A well was characterized by a remarkable and constant increase in salinity, which became almost fivefold, increasing from 3800 to 17000 mg/l, between 1976 and 1985. In the same period the gas concentration decreased by only 10%, while various gases internal ratios remained unchanged. The behavior in the liquid phase was different. In this phase the variations of some internal ratios of the major compounds were significant. This can be explained only by a gradual increase of re-equilibrated seawater inflow.

The available data for the period 1985-1989 (unpublished) show that starting from 1986 this seawater inflow became constant. During the whole production history, however, the enthalpy of the produced fluid remained unchanged.

With reference to the results from the wells drilled, some authors have suggested that the transverse offset in the LERZ can condition the distribution of permeability and of temperatures in depth. There is not enough evidence, however, of this transverse offset nor of its effect on the characteristics of the geothermal area drilled. In fact, the geophysic data available outline only a transverse disturbance zone whose structural significance is uncertain and whose width is not well-defined. Furthermore, all the other wells drilled, apart from the Ashida 1, fall into this zone of transverse disturbance whether they are productive or not.

7. CONCLUSIONS AND RECOMMENDATIONS

The purpose of this study can be summarized as follows:

- evaluation of studies and geothermal explorations carried out to date in the KERZ area;
- identification of future activities to be carried out in order to verify and characterize the geothermal resource before starting a 600 MW development project.

This study has also allowed the preparation of a report which is an updated collection and summary of available data with the aim of obtaining a geothermal characterization of the Kilauea East Rift Zone and more in particular, of the area selected for the project (GRS).

7.1. CONCLUSIONS

The geological and geophysical data make possible a reliable structural modelling of the rift which has a width of 4-5 km at surface and of 12-19 km at depth. Up to a depth of about 6 km there is a Dike Complex which is linked to the central magmatic chamber of the Kilauea. Deeper down, the Deep Rift System is formed of intrusions linked to a second and deeper magmatic chamber.

A widespread geothermal anomaly can be hypothesized in the entire KERZ area and can be associated with the deep intrusions such as the Dike Complex and the satellite magmatic chambers.

Surface surveys performed specifically for geothermal exploration do not give significant information about the presence of deep geothermal reservoirs. In fact, most of the geoelectric surveys available, besides being dishomogeneous and concentrated in the LERZ zone, are characterized only by little depth of investigation.

Exploratory wells drilled to date are located in a rather small area compared to the one selected for the project (GRS). These wells have confirmed the existence of a high temperature hydrothermal system. A tentative model of this system has been made on the basis of few data regarding mineralization, temperatures and electric resistivity. The reservoir is mainly hosted in the Epidote and Actinolite Zone and in a smaller extent in the Chlorite Zone. The Montmorillonite Zone acts as a geothermal cover and separates the deep reservoir from the shallow aquifer which is hosted in the slightly altered stratigraphic interval.

These processes of mineralization and hydrothermal alteration should be rather widespread in the rift.

Because the area investigated by the wells is very small, the model cannot be inferred with reliability to the entire area of the project (GRS).

The wells have confirmed a high thermal anomaly and have shown a discontinuous permeability due to fracture systems mainly developed lengthwise the rift. In a transverse direction the low permeability is reduced by the presence of dikes. The fracture system can be closed by self-sealing but can recur due to the continuous volcanotectonic activity.

Fluids produced by the wells HGP-A, KS-1A and KS-2 probably come from the same water-dominated reservoir or from reservoirs with similar thermal characteristics, even though interference tests have not been carried out to verify these hypotheses. The salinity of the fluid produced by the HGP-A well has increased from 3800 to 17000 mg/l while its thermal characteristics have remained unaltered. Chemical evidence lead to the hypothesis that a seawater origin inflow occurs in some peripheral parts of the reservoir. These phenomena could be better investigated if suitable isotopic data were available.

The high amount of rainfall, quickly drained by a shallow aquifer, and the depth of the water table overlap any surface manifestations of the deep reservoir.

In the rift this aquifer, whose sea level temperature ranges between 30 and 55°C, causes a certain heat leaching. Great quantities of warm waters discharge, in fact, lengthwise and in a lesser extent through the southern flank. Locally spots of high temperature water points can be due to deep fluid uprising along fractures or to heating by a recently intruded dike.

Only the fluid of the GTW-3 can be surely related to a deep geothermal component. The other surface samples are mainly diluted fresh waters or fresh waters mixed with a different percentage of seawater. In the latter, deep contamination, even if present, cannot be detected by the available chemical analyses.

Wells drilled in the southern edge of the rift revealed temperatures lower than those inside the rift at the same depth. However, temperature values significant for an industrial exploitation can be expected also in the southern area at a slightly higher depth (2000-2500m). High seismicity near the southern flank might indicate a fractured system with an hydrothermal circulation. Therefore, the possibility of enlarging GRS southward could be consistent with the targets of the project. A proper evaluation of the seawater inflow and of the volcanic hazard due to the lava flows in this area must be carefully performed before starting the deep exploratory program.

The northern flank of the LERZ can also be indicated as a potential enlargement area for the GRS. In fact, since the LERZ is active on both flanks, it may be subject to intrusion and fracturing phenomena in its northern part, too. It can also be hypothesized that the thermal anomaly in the northern part of the LERZ is affected by the convergence both of the Mauna Loa and of the Kilauea East Rift.

7.2. RECOMMENDATIONS

It is clear that, before starting a development project which requires a large capital investment, it is necessary to perform many activities in order to verify the presence of a geothermal resource industrially exploitable, and to estimate the field potential.

The activities so far carried out are not sufficient and have been neither coordinated nor finalized in a common criteria of geothermal characterization of the area. Data and information collected are therefore dishomogeneous and cannot guarantee an adequate interpretation.

As a consequence, the following recommendations aimed at the verification and characterization of the resource, are related both to the technical and management aspects.

7.2.1. TECHNICAL RECOMMENDATIONS

Although the presence of a geothermal anomaly in the entire KERZ is reliable, the presence and the characteristics of an industrially exploitable geothermal reservoir will have to be still verified. With this target in mind, the drilling of 10 to 15 deep exploratory wells and measurements and tests for the reservoir and fluids characterization is a priority program considering the KERZ structure.

Data from these wells will be analyzed and interpreted together with all existing information and data from SOH, at present being drilled. In November 1989 ENEL recommended to carry out a program of measurements, logs and tests for the SOH. In particular:

- temperature and pressure surveys in static conditions;
- injection test for the fractured layers localization and for the hydraulic characteristics determination;
- physical measurements on some cores of the main lithological units (bulk and grain density, porosity, magnetic susceptibility, thermal conductivity and specific heat) and relative paragenesis and mineralization analyses;

- open hole geophysical well-loggings (Dual Laterlog and/or Dual Induction Log, Formation Density Compensated Log and, at least in one well, Sonic Log);
- production tests with fluid samplings for chemical and isotopic determinations.

As regards the drilling of the deep exploratory wells, the first 4 to 5 wells might be appropriately located in order to give information about the entire GRS area.

For these wells it is advisable to:

- collect systematically cuttings and withdrawal of cores of the main lithological and/or mineralized units. These samples will be used to define the well geology (stratigraphy, alteration degree, hydrothermal paragenesis, etc.). The cores can also be used for physical measurements (bulk and grain density, porosity, magnetic susceptibility, thermal conductivity);
- perform geophysical logs in order to obtain a direct characterization of the formation resistivity (Dual Lateral Log and/or Dual Induction Log), of the porosity (Compensated Neutron Log), of the density (Formation Density Compensated) and the propagation velocity of the P waves (Borehole Compensated Sonic Log);
- perform temperature recovery survey at various depths, during drilling, whose extrapolation will indicate the true formation temperature;
- perform temperature and pressure surveys in static and dynamic conditions (injection and production) to get information about the fractured layers;
- perform injection, flow and interference tests;
- collect fluid samples during the flow tests for chemical and isotopic analyses.

The data collected in these first exploratory wells will be also used in order to evaluate the opportunity of performing some geophysical surface surveys to better characterize the area and hence to better locate the subsequent deep exploratory wells.

As regards the other wells, a reduced program of geophysical logs could be foreseen according to the results from the previous ones.

According to the structure of the KERZ the directional drilling in a transverse direction of the rift should be considered. In this way the probability of crossing the subvertical fracture system is increased and, therefore, the probability of encountering production layers is increased as well.

Due to the presence of high temperature and low permeability, water or air drilling rather than mud drilling is advisable to avoid fracture plugging.

The data from all the exploratory wells will also be used for the determination of the parameters necessary for planning the development project:

- depth of the wells;
- success ratio of the wells;
- mean specific productivity for each productive well (MW/well);
- enthalpy of the produced fluid in order to plan the number of reinjection wells;
- physical-chemical characteristics of the fluids produced with particular reference to corrosion and scaling aspects.

Given the target of the project, the enlargement of the GRS should be considered as reported in paragraph 7.1.

After having drilled some deep exploratory wells within the GRS and having verified the presence of a geothermal resource, exploratory drillings outside the GRS could give indication about areas in which the enlargement may be planned.

However, the exploratory wells and the surface supplementary surveys cannot guarantee the assessment of the field potential, unless development projects start to operate in some areas with the installation of the first power plants and with the fluid production. Data collection during this first periods of exploitation is very important for the field potential evaluation. According to the target of the project (600 MW of installed capacity), a step by step development strategy is therefore recommended.

At present, two private companies (ORMAT and TRUE/MID PACIFIC) are operating with their own mining leases in the project area. ORMAT will drill wells for the development of the well-known Kapoho area in which power plants for a total of 30 MW (gross) will be installed. TMP is drilling some deep wells in the Kilauea Middle East Rift in order to verify the presence of geothermal resource in this area too.

The planning and the execution of a systematic program of measurements and tests for the wells drilled by these two Companies could allow to collect data and information very useful for the resource verification.

Some of these wells could be considered as a part of the deep drilling exploratory program recommended above.

Moreover, a systematic program of measurements and tests (interference tests, in particular) during the ORMAT field exploitation will be very important for the field potential evaluation.

Some surface studies and surveys, integrating those already made, are be carried out together with the drilling of the exploratory wells. These surveys should be calibrated with the well data to assure a much better characterization of the area.

The printing of an up-dated geological map (on the basis of the existing data) with a scale of 1:50000 is particularly advisable for geology.

As far as the surface supplementary geophysical surveys are concerned, the boundaries of the Dike Complex and of the structural reconstruction on a regional scale (already adequately defined) would be precisely defined by a detailed gravimetric survey. Therefore, before performing a new survey, it is advisable to ascertain whether local research organizations (e.g. Hawaiian Volcano Observatory) have carried out gravimetric measurements in the KERZ which are more recent than the officially known data (Kinoshita, 1965). If so, it is advisable to initiate the printing of a more up-to-date map of the Bouguer anomalies. Density measurements on significant well samples could improve the interpretation of the data.

The magnetometric data available, after the 1978 low altitude survey (Flanigan and others), can be considered sufficient, but a better interpretation is recommended. Measurements of the magnetic susceptibility on surface and well samples are necessary to improve this interpretation.

The very high seismicity which characterizes the KERZ and the South Flank makes it difficult to use the seismic data as an indicator of permeable areas due to fractures. In order to increase the accuracy of the hypocenter determination and of the focal mechanism solutions, an increase in the seismometric stations number should be necessary as well the location of some stations inside wells, but times and costs of such a survey seem to be too long and to high, respectively. However, the particular concentration of seismic events in the South Flank indicate an existing tectonic activity which could be linked to intense fracturing and, therefore, to the presence of a hydrothermal system. If an enlargement of the selected zone should be necessary for the project, it would be beneficial to verify this hypothesis with the drilling of a deep exploratory well.

Regarding results obtained in the Kapoho area with the drilling of deep exploratory wells and the elaborated geothermal model, it is believed that electric surface surveys can make possible the characterization of the layers. These layers, with their different degrees of alteration, play different roles in the geothermal system.

Should the future exploratory wells confirm the possibility to differentiate the layers from a geoelectrical point of view, the performing a surface survey as homogeneous as possible over the entire KERZ and adjacent zones, is recommended. Because a reconstruction of the electric characteristics in a large range of depths is necessary, an integrated electromagnetic and magnetotelluric survey is advisable. Calibration Vertical Electric Soundings (VES) with AB spacing of at least 6-8 km is also recommended close to some wells.

Regarding <u>geochemical surveys</u> a re-sampling of warm and cold water points (see Fig.1/An.B) is recommended in the Puna area for chemical and isotopic analyses. Calculations of pH, alkalinity and H_2S should be carried out. Besides the calculations of main components and of SiO_2, the calculations of at least Li, Rb, Cs, Br, NH_3 and As should be carried out in the laboratory.

7.2.2. MANAGEMENT RECOMMENDATIONS

Given the extent of the Geothermal/Cable Project, which includes the installation of a 600 MW capacity, management and coordination of the various activities related to the exploration, to the development and to the field exploitation must be dealt with by a single Organization.

The geothermoelectric production is characterized by large capital investment and low operating costs. Therefore a reliable geothermal resource assessment is very important before starting with a development project. The planning and coordination of all the exploration activities is essential together with the integrated interpretation of all data collected for the resource characterization and assessment of over the entire project area.

The Organization will have to:

- plan the activities necessary for adequate resource verification;
- check and guarantee that a coordinated program of surface and deep exploration are carried out in the entire area of the project and that all the measurements, tests, samplings

and analyses are carried out using similar methodologies;

- deal with an integrated interpretation of all the information obtained in order to up-date the geothermal model of the area;
- assure the flexibility of these programs according to the collection of new data;
- finalize a budget of State supports necessary for integrating the above-mentioned activities;
- plan a project development program on the basis of all the information obtained with the activities related to the resource verification and characterization.

At present, three different operators are working independently in the project area: University of Hawaii, TMP and ORMAT. These last two are private Companies working with their own mining leases in areas located within the Geothermal Selected Subzones. The coordination of the activities of the three operators, especially for the planning and execution of measurements, surveys and tests, is very important at this stage of the project to collect significative data and information.

An Organization charged with the coordination of all the activities, with geothermal experts both for management and for the technical-scientific side, is therefore an absolute priority for this project. The Organization should deal with relations between public and private bodies and co-ordinate all surface and deep exploration activities.

A single global management is also important during the subsequent stages of project development and field exploitation: the overall knowledge of the geothermal system is fundamental for the planning of development and of exploitation strategies (production and reinjection) aimed at optimizing the exploitation of the geothermal resource.

The subdivision of the project area between different operators, which work independently, does not permit an overall knowledge of the geothermal field and therefore it does not permit a plausible planning of the optimum capacity to be installed for a 20-30 years service. Each operator may have its own production and reinjection strategy beyond an overall management strategy for the resource exploitation.

The Geysers geothermal field experience confirms the above mentioned problems, while in Italy, where ENEL is the only owner and operator in the geothermal fields under exploitation, the overall knowledge of these systems has allowed to plan management strategies in order to optimize the geothermal resource exploitation.